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VOLCANIC CENTERS OF THE SUNLIGHT AREA PARK COUNTY, WYOMING[1]

WILLARD H. PARSONS
Hamilton College

ABSTRACT

The Sunlight area is located in the northern Absaroka Mountains of northwestern Wyoming. The rocks are dominantly Tertiary pyroclastics and basalt flows which rest on flat-lying Paleozoic limestones. The pyroclastics were thrown out from a large number of vents in explosive eruptions and were later modified by mud flows. Two major volcanic centers and one small explosive vent have been found in the Sunlight area. Intrusions of stocks, laccoliths, plugs, cone sheets, and radial dike systems are closely related to the volcanic centers. The radial dikes were formed by magma moving horizontally outward from the intrusive centers in vertical cracks. The Sunlight centers are similar in many respects to the Scottish Tertiary volcanoes. The rock types in general show a normal differentiation series from olivine gabbro and basalt through diorite and andesite to sodic syenite and trachyte. Several unusual orthoclase-bearing basalts, known as the absarokite series, are also present.

INTRODUCTION

The Sunlight area is located in the northern Absaroka Mountains in the Shoshone National Forest, Park County, Wyoming (mostly included in the Crandall Sheet of the Absaroka Folio). The area studied trends northeast–southwest along Sunlight Creek and is about 20 by 8 miles in extent. It was mapped first by Arnold Hague in 1893 and 1897.[2] The small sulphur deposits occurring within the area have been described by Hewett.[3]

[1] Presented in part at the Forty-ninth Annual Meeting of the Geological Society of America in Cincinnati, Ohio, December, 1936. W. H. Parsons, "Volcanic Centers of the Sunlight Area, Wyoming" (abstract), *Geol. Soc. Amer. Proc. 1936* (1937), pp. 93–94.

[2] Arnold Hague, Absaroka Folio, No. 52, *U.S. Geol. Surv. Atlas* (1899).

[3] D. F. Hewett, "The Sulphur Deposits of Sunlight Basin, Wyoming," *U.S. Geol. Surv. Bull. 530* (1911), pp. 350–62.

The field work for the present study was done during the summers of 1934 and 1935 and represents one of the co-operative studies of the Yellowstone–Bighorn Research Association. Vertical areal photographs, taken for the United States Forest Service, were used in making the geologic maps.

GENERAL GEOLOGIC SETTING

The area lies within the Absaroka Mountains, which are a part of the Middle Rocky Mountain physiographic province.[4] It is characterized by rugged mountain topography, well dissected, and modified by mountain glaciation. The relief varies between 4,500 feet in the western part and 2,000 feet in the eastern section. The latter section, called Sunlight Basin, has a very broad and flat valley which was a sediment-filled glacial lake.[5]

The rocks of the region consist essentially of flat-lying Paleozoic limestones overlain by extrusive Tertiary pyroclastics and lava flows. These are cut by Tertiary intrusive rocks in the form of small stocks, plugs, and laccoliths grouped in two main centers, and hundreds of dikes most of which radiate from the two centers. These intrusive centers represent major volcanic centers from which some of the pyroclastics and interbedded flows were extruded.

PALEOZOIC ROCKS

The Paleozoic rocks, totaling about 15 square miles of outcrop area, are exposed in the eastern part of the area (see Fig. 1). These range in age from Cambrian to Mississippian, making up a total section about 2,000 feet thick. The various formations present and their approximate thicknesses are as follows:[6] Middle Cambrian, Gros Ventre shales, 75–95 feet; Upper Cambrian, Gallatin, or Pilgrim limestone, 75 feet; Grove Creek shales and limestones, 150 feet; and an unnamed formation of yellowish sandy shales with

[4] N. M. Fenneman, *Physiography of the Western United States* (New York: McGraw-Hill Book Co., 1931), map.

[5] Parsons, "Glaciation in the Sunlight Valley, Wyoming" (abstract), *Geol. Soc. Amer. Proc., 1937* (1938); forthcoming article on the glacial features of the Sunlight area, Wyoming.

[6] For a more detailed description of the stratigraphy, paleontology, and structure of these formations, see Parsons, "The Geology and Ore Deposits of the Sunlight Area, Park County, Wyoming" (Princeton University Doctor's dissertation [1936]).

graptolites, 10–20 feet; Ordovician, Bighorn dolomite, 270 feet; Devonian, Jefferson dolomite, 280–90 feet; and Three Forks sandy shales and thin limestones, 100 feet; Mississippian, Madison limestone, 800–1,000 feet. Just east of the area mapped, the section continues uninterrupted down through the Flathead quartzite into the pre-Cambrian granite of the Clarks Fork region. These Paleozoic rocks are practically flat-lying, though a few local folds and minor faults are present, especially in the vicinity of Elk Creek.

PREVOLCANIC TOPOGRAPHY

Prior to the extrusion of the first volcanic rocks a surface of considerable relief had been produced by erosion of the Paleozoic rocks. This unconformity is visible in many parts of the Absaroka Mountains.[7] In the Sunlight area this prevolcanic surface had a relief of 1,000–1,500 feet and was characterized by high, flat plateaus or mesas with steep sides and wide, flat-bottomed valleys. One high plateau existed in the area between Elk Creek, White Mountain, and Russell Peak, while the remnants of another are seen today in Little Bald Ridge and the limestone outcrops south of it. These relationships may be seen in Figures 1 and 2. The limestone-volcanic contact has an elevation of about 8,000 feet in the Elk Creek region and between 8,000 and 8,500 feet on Little Bald Ridge. This same contact is just below 7,000 feet today in the 3–4-mile-wide valley between these two ancient plateaus. This prevolcanic valley trends at right angles to the present Sunlight Creek Valley.

EXTRUSIVE ROCKS

The volcanic rocks of the Absaroka Range were divided into the following six groups by Hague:[8] early acid breccia, early basic breccia, early basalt sheets, late acid breccia, late basic breccia, and late basalt sheets. In local areas later workers[9] have subdivided the early basic breccia into an early basic tuff and early basic breccia. In the Sunlight area the following were recognized: early basic

[7] John T. Rouse, "Genesis and Structural Relationships of the Absaroka Volcanic Rocks, Wyoming," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), p. 1261.

[8] *Op. cit.*

[9] Rouse, "The Volcanic Rocks of the Valley Area, Park County, Wyoming," *Trans. Amer. Geophy. Union 16th Ann. Meeting* (1935), pp. 274–84.

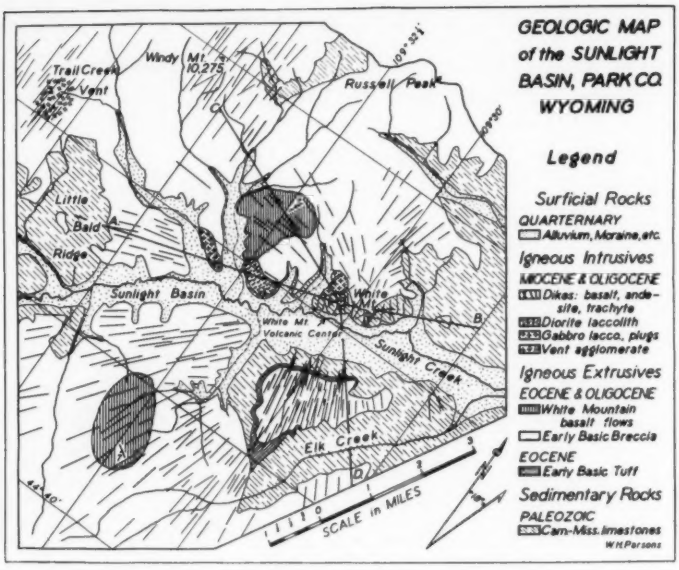
GEOLOGIC MAP
of the SUNLIGHT
BASIN, PARK CO.
WYOMING
Legend
Surficial Rocks
QUARTERNARY
Alluvium, Moraine, etc.
Igneous Intrusives
MIOCENE & OLIGOCENE
Dikes: basalt, andesite, trachyte
Diorite laccolith
Gabbro lacco., plugs
Vent agglomerate
Igneous Extrusives
EOCENE & OLIGOCENE
White Mountain basalt flows
Early Basic Breccia
EOCENE
Early Basic Tuff
Sedimentary Rocks
PALEOZOIC
Cam-Miss. limestones
W.H. Parsons
Trail Creek Vent
Windy Mt. 10,275'
Russell Peak
Little Bald Ridge
Sunlight Basin
White Mt. Volcanic Center
Sunlight Creek
Elk Creek
SCALE in MILES
A
B
C
D

FIG. 1

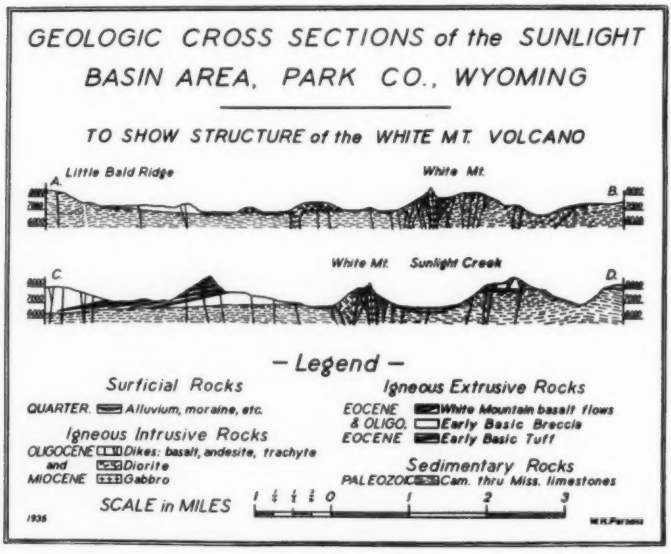
GEOLOGIC CROSS SECTIONS of the SUNLIGHT
BASIN AREA, PARK CO., WYOMING
TO SHOW STRUCTURE of the WHITE MT. VOLCANO
Little Bald Ridge
White Mt.
A
B
C
D
White Mt.
Sunlight Creek
— Legend —
Surficial Rocks
QUARTER. Alluvium, moraine, etc.
Igneous Intrusive Rocks
OLIGOCENE and MIOCENE
Dikes: basalt, andesite, trachyte
Diorite
Gabbro
Igneous Extrusive Rocks
EOCENE & OLIGO.
White Mountain basalt flows
Early Basic Breccia
EOCENE
Early Basic Tuff
Sedimentary Rocks
PALEOZOIC
Cam. thru Miss. limestones
SCALE in MILES
0
1
2
3
1935
W.H. Parsons

FIG. 2

tuff, early basic breccia, and early basalt sheets. The late series, which outcrops immediately southwest of the Sunlight area, has been entirely removed by erosion in this area, if, indeed, it was ever present.

EARLY BASIC TUFF

Locally at the base of Hague's early basic breccia series a light-colored andesitic tuff is present. Near Valley, Wyoming, this tuff is 400–900 feet thick and, in mapping, has been separated from the early basic breccia.[10] In the Sunlight area most of this early basic tuff occurs on the mountain between Elk and Sunlight creeks and rests on the Madison limestone. The average thickness of the tuffs is 200 feet with a maximum of 300 feet. The series is well bedded, variegated in color and texture, and is essentially flat-lying, though local dips up to 10° were recognized.

The tuffs are fine- to medium-grained with occasional interbedded volcanic conglomerates. They are well consolidated, very compact, and resistant to erosion, weathering into steep slopes. The series is light in color, generally a brownish- or grayish-white, although purple, blue, or green beds do occur.

The tuffs are surprisingly similar in composition in spite of changes in color and grain size. They are crystal tuffs composed chiefly of angular fragments of euhedral crystals with a few fragments of rock in a very fine-grained to aphanitic groundmass of ash particles. The crystal fragments are plagioclase, augite, biotite, hornblende, magnetite, and apatite. Plagioclase, either labradorite or andesine, is the most abundant. Some of the thinner tuff beds contain fossil leaves and fragments of plant stems. The volcanic conglomerates, occurring as thin lenses, contain pebbles ($\frac{1}{4}$–$\frac{1}{2}$ inch) of andesite and basalt porphyry in a groundmass of tuffaceous material. Pebbles of pre-Cambrian granite and Paleozoic limestone are present in some beds. The composition of the groundmass of the conglomerates is similar to that of the crystal tuffs.

After the deposition of these tuffs there was some slight deformation followed by erosion which completely removed the tuffs from much of the area.

[10] *Ibid.*, p. 276.

EARLY BASIC BRECCIA

This unit is composed of interbedded volcanic breccias, volcanic conglomerates, and rare agglomerates.[11] This series outcrops over approximately three-fourths of the area and has a maximum thickness of 5,000 feet. In appearance and texture the breccias vary considerably. The bulk of the series is composed of dark-brown volcanic breccias and conglomerates with angular to subangular fragments

FIG. 3—A coarse agglomerate near the Trail Creek vent with large rounded fragments typical of the agglomerates.

an inch or two in diameter in a matrix of small rock fragments, crystals, fine ash, and glass dust. Locally 6–10-foot fragments are numerous and 100–200-foot masses are known. Great variation occurs in the degree of compaction and in the type of weathering. Some parts of the breccia series are very compact and stony while others are very crumbly and scoriaceous.

Coarse agglomerates are fairly common in a few places with subrounded fragments of very varied size but averaging a foot or two

[11] The terminology of the pyroclastic rocks suggested by C. K. Wentworth and Howell Williams ("The Classification and Terminology of the Pyroclastic Rocks," *Nat. Res. Counc. Bull. 89* [1932], pp. 45–51) is being used throughout this paper.

in diameter. One of these may be seen in Figure 3. Their groundmass, which is composed of small fragments and fine ash, does not have rounded material in it.

The early basic breccia series in the Sunlight area does not have the well-defined bedding characteristic of this unit in other parts of the Absarokas because of the high proportion of volcanic breccias present in the Sunlight as against volcanic conglomerates elsewhere. Most of this series shows little or no bedding. A well-bedded structure is observed only where volcanic conglomerate beds are common, and such bedding is approximately horizontal. Well-defined and steeply dipping bedding also occurs locally with interbedded flows or tuffs in association with actual vents. These vent structures will be described later.

The composition of these volcanic breccias, volcanic conglomerates, and agglomerates is very similar. The fragments, which are all porphyritic, are mainly andesites, with some basalts, olivine basalts, and rarely hornblende andesites and trachyandesites. The common minerals in the fragments and groundmass are labradorite, andesine, and augite, with some hornblende, biotite, and magnetite. Olivine is rare and usually altered to iddingsite or serpentine. Some of these primary minerals have been altered, principally in the Sunlight mining region,[12] to chlorite, sericite, calcite, and epidote.

EARLY BASALT SHEETS

The early basalt sheets, resting unconformably on the early basic breccia, comprise a series, 750–975 feet thick in the southern Absarokas, of many individual porphyritic basalt flows each 50–100 feet thick. Rouse[13] has divided the series into six groups on the basis of stratigraphic position and petrographic character.

In the Sunlight area a few small remnants of the early basalt sheets occur on the tops of some of the higher peaks, where they weather into cliffs. Such remnants are illustrated in Figure 4, and the areal extent of the sheets is shown in Figure 7. All the flows are porphyritic, some having labradorite phenocrysts and others having only augite and olivine phenocrysts. The groundmass in either case

[12] Parsons, "The Ore Deposits of the Sunlight Mining Region, Park County, Wyoming," *Econ. Geol.*, Vol. XXXII (1937), pp. 844–46.

[13] "The Volcanic Rocks of the Valley Area . . . . ," *op. cit.*, pp. 274–84.

contains small crystals of magnetite, augite, and tiny labradorite laths with a minor amount of interstitial glass.

LIMESTONE XENOLITHS IN THE BRECCIAS

Fragments of Paleozoic limestone and pre-Cambrian granite occur in the early basic breccia series at various places. A few beds, especially some of the volcanic conglomerates, contain many small rounded pebbles of limestone and granite of the same size as the

FIG. 4.—One of the mountain peaks at the head of Jaggar Creek capped by beds of early basalt sheets. The less steep slopes are composed of unbedded early basic breccia. The dikes cutting the breccias are part of the Sunlight radial system.

volcanic fragments. Single limestone and granite fragments, 1–4 feet in diameter, occur scattered through the breccia series. Such pebbles and fragments appear to have been thrown from vents during explosive eruptions together with the volcanic material.

Some of the limestone blocks, however, have reached their position in a different manner. Such angular blocks are 10–100 feet in length (see Fig. 5), and one is 300 feet. In an area of a few acres, ten or twelve of these blocks may be present, and the breccia in which the blocks are contained is composed of small volcanic fragments (1–2 inches). The limestone fragments have not been metamorphosed and are not fractured or shattered.

One group of such limestone xenoliths is embedded in a very coarse vent agglomerate in close proximity to the Trail Creek vent.

These blocks of limestone were torn off deep within the throat of the volcano and carried to the surface without being broken or actually thrown through the air.

The other groups of large blocks do not occur near vents. They are embedded in the volcanic breccias which fill the prevolcanic valleys near and below the edges of the old limestone mesas. Such blocks were evidently emplaced by mud flows. Volcanic breccias

FIG. 5.—A large limestone block in the early basic breccia. This block is about 100 feet long.

covered the prevolcanic mesas, moved down as mud flows into the old valleys, and picked up or tore loose fragments of the country rock (limestone) as they moved over the edges of the mesas.

AGE AND ORIGIN OF THE EXTRUSIVES

The age of the Absaroka volcanic rocks has yet to be decided. The subject is well summed up by Rouse,[14] who favors an Eocene age for the volcanics of the Absaroka volcanic field with a possibility that activity continued into the Oligocene epoch. The present writer accepts this age determination with the added suggestion that intrusive activity continued into the Miocene epoch.

The early basic breccia series reached the surface and was dispersed over large areas through various types of volcanic eruptions.

[14] "Genesis and Structural Relationships . . . . ," *op. cit.*, pp. 1271–72.

These include vulcanian eruptions, nuées ardentes, flows, and dry avalanches. In some cases the shape and form of these eruptions were subsequently modified by mud flows and stream action. Vulcanian eruptions were important from many centers, though the material so emplaced was soon modified and shifted by stream erosion and dry avalanches. However, mud flows were responsible for most of the modification and destroyed all original bedding, moved material away from the vents, and produced a fairly level surface on the breccias. It has been shown that mud flows take place on 1° or 2° slopes for great distances and are common phenomena in connection with vulcanism.[15]

Peléan eruptions or nuées ardentes may have been present, as the eruptions in this region were decidedly violent, a fact evidenced by the very coarse explosion breccias sometimes seen. In such violent types of eruptions, nueés ardentes are to be expected especially in andesitic lavas.

A few flow breccias are associated with basalt lava flows near the White Mountain vent. This type of eruption, however, was of little importance in this area. The fragments of these few flow breccias are embedded in a crystalline or glassy groundmass.

The pyroclastics were erupted from a great many small vents and a few medium-sized volcanoes scattered over the Absarokas. The ejectamenta from these numerous vents have been so intermixed that it is usually impossible to locate the vent from which particular breccias came. Two medium-sized volcanoes and one smaller one have been recognized in the Sunlight area from which some of the local breccias and flows came. The early basalt sheets were extruded as fissure eruptions.

For a discussion of the problems of origin of the volcanic rocks the reader is referred to Rouse's recent publication.[16]

VOLCANIC CENTERS AND INTRUSIVE BODIES

In the Sunlight area two good-sized volcanoes and one small vent have been located. The criteria used to recognize old volcanoes are

[15] C. A. Anderson, "The Tucson Formation of Northern California with a Discussion of the Origin of Volcanic Breccias," *Univ. Calif. Pub., Dept. Geol. Sci. Bull.*, Vol. XXIII (1933), pp. 215–76.

[16] "Genesis and Structural Relationships . . . . ," *op. cit.*, pp. 1273–82.

the following: presence of plugs or volcanic necks, intense contact metamorphism around the plug, coarse vent agglomerates, initial dips in the vicinity of a vent, radial dike systems, and centrally located groups of other small intrusive bodies.

WHITE MOUNTAIN VOLCANO

White Mountain (Figs. 1 and 6) represents one such volcanic and intrusive center. The central plug (Fig. 6) is a nearly vertical,

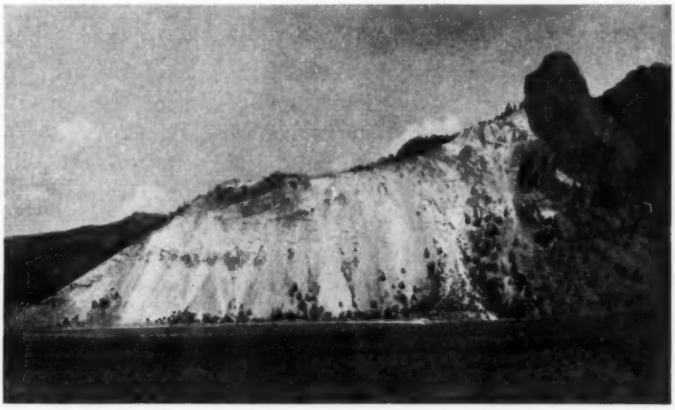

FIG. 6.—The southwest spur of White Mountain showing the white metamorphosed limestone and the neck of the White Mountain volcano on the right. Part of the network of branching dikes and sills may be seen just left of the plug.

elliptical, composite pipe, 1,000 by 500 feet in size, with vertical sheetlike jointing. It is composed of rocks varying from augite diorite to olivine gabbro which cut one another in an irregular fashion indicating successive periods of intrusion. Very near this central neck is a flaring, funnel-like stock or plug of gabbro about a half-mile in diameter.

Just west of this center are two domelike laccoliths of augite diorite and olivine gabbro, respectively, resting on unmetamorphosed limestone and having their upper, steeply dipping contacts concordant with the overlying volcanic breccias.

The limestone of White Mountain is cut by a large number of narrow, irregular dikes and sills which form an anastomosing net-

work. In addition, hundreds of vertical dikes of andesite porphyry, averaging 6 feet in width, radiate 2–6 miles from White Mountain. They are far more numerous in the volcanic breccias than in the underlying limestones, radiating farther from the center to the south and west where breccias outcrop than to the north and east where limestones outcrop. It is apparent that the volcanic breccias were less resistant to intrusion than the limestones.

The limestone of White Mountain is entirely recrystallized by contact metamorphism to form a coarse-grained marble. This white marble is in strong contrast to the usual gray-white of the unaltered Paleozoic limestones. Most of the marble is rather pure calcite and dolomite, but a small amount is impure, fine-grained, dense, and of variegated color. The more common metamorphic minerals, which occur in small veins and disseminations, are serpentine, diopside, wollastonite, and epidote. Iron oxides, olivine, vesuvianite, garnet, phlogopite, and spinel are present in much smaller amounts. The volcanic breccias overlying the limestone near the plug have been indurated and, though unchanged in composition, are in striking contrast to the unaltered breccias.

Away from White Mountain, the limestones and breccias, even adjacent to the large laccoliths, are unmetamorphosed. The mere intrusion of a small plug, therefore, was not sufficient to produce the metamorphism of White Mountain, but heat and solutions must have been applied over a long period of time, as is possible only in the walls of the conduit of a volcano active for centuries.

Thinly bedded lava flows and volcanic breccias covering two areas, each comprising 1–2 square miles, are believed to have been extruded from this volcano. They lie within the early basic breccia series, one about 2 miles west, the other 4–5 miles southwest of White Mountain. They consist of interbedded vesicular basalt lava flows, averaging 10 feet thick, and scoriaceous volcanic breccias, some of which are actual flow breccias. These beds all dip away from the volcano at an angle of about 7°.

It is difficult to state how much of the early basic breccia was erupted from this vent. The composition, position, and attitude of the lava beds just described indicates that they were poured from this center. The plug rocks, lava flows, and flow breccias from this

vent are very basic in composition, more so than the majority of the early basic breccias. This might indicate that much of the early basic breccias in the Sunlight area was erupted from other vents. One flow area is near the base of the early basic breccias, the other more than halfway up in the 5,000-foot series, showing that the volcano was active during a considerable part of the period of deposition of the early basic breccias. A rough idea of the size of the volcano is obtained by projecting the dips of the lava flows to White Mountain. Such a volcano would have a radius of approximately 5 miles and a height of 5,000 feet.

A 10-foot bed of tuff outcropping on one side of White Mountain appears unusual. Three-inch inclusions of limestone metamorphosed on their edges indicate that this tuff was deposited at a very high temperature and is perhaps a deposit similar to the great sandflow of the Valley of Ten Thousand Smokes,[17] which Fenner believes was deposited by a nuée ardente. Possibly, therefore, the early eruptions of the White Mountain volcano included nuées ardentes.

TRAIL CREEK VENT

This vent is located southwest of Windy Mountain (see Fig. 1). It is approximately 1,000 feet in diameter and is filled with a very coarse vent agglomerate, which in turn is cut by a few irregular, altered dikes. The vent agglomerate is composed of large subrounded fragments, 8–15 feet in diameter, in a matrix of fine ash and smaller fragments, with occasional limestone blocks 100 feet long. As the limestone is stratigraphically 1,000 feet below the surface here, these blocks represent xenoliths torn from the walls of the conduit by the ascending agglomerate. The vent agglomerate has the same composition as the early basic breccia series. Agglomerates with smaller fragments and volcanic breccias occurring for a mile or two on all sides of the vent definitely were thrown out from this center. The Trail Creek Volcano represents one of many small vents which poured out a majority of the early basic breccias. The life of such volcanoes was rather short but violent.

There is a north–south dike swarm associated with the vent and a similar parallel swarm 3 miles northeast on Windy Mountain.

[17] C. N. Fenner, "The Origin and Mode of Emplacement of the Great Tuff Deposit of the Valley of Ten Thousand Smokes," *Geophy. Lab. Paper 480* (1923).

These are vertical dikes similar in size and composition to the White Mountain radial swarm but strike at right angles to the latter.

SUNLIGHT VOLCANO

A volcano is located in the western part of the area in the high mountains between branches of Sunlight Creek and the headwaters of the north fork of the Shoshone River (Fig. 7). Here are found at

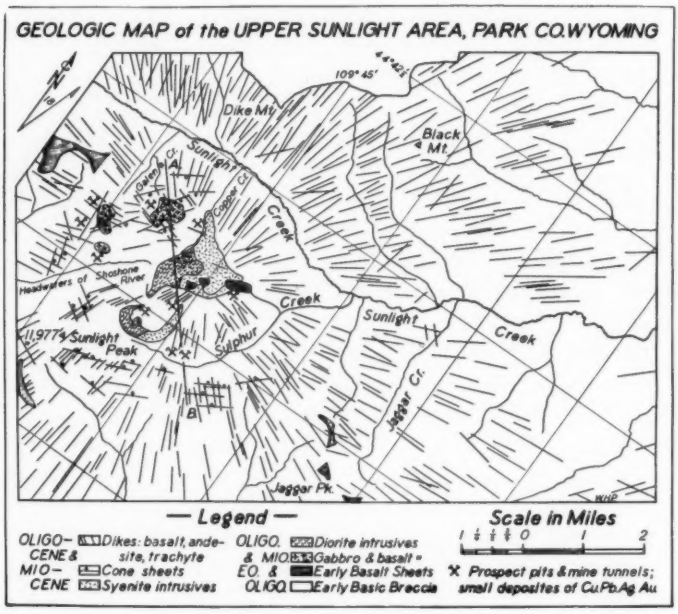

FIG. 7

least two actual vents, a small stock, a ring dike, cone sheets, and a radial dike system.

A denuded volcanic cone at the head of Galena Creek has a central plug, composed of basalt porphyry, about 3,000 by 2,000 feet in length and width. Surrounding the plug, interbedded tuffs, volcanic breccias, and andesitic lava flows dip as much as 20°–30° away from the center (Fig. 8). Another smaller but similar plug filled with basalt porphyry and intrusive breccia is located a mile to the south-

west. Around it the volcanic breccias are much disturbed, faulted, and broken, and great blocks of the breccia rest at all angles at the edge of the vent. A steep-sided cone of pyroclastic material and lava was built up around the Galena Creek vent. The smaller plug fills a parasitic vent on the side of the Galena Creek cone, through which a very violent explosion took place, causing widespread disruption and faulting of the beds.

A small, triangular, syenitic stock, covering approximately 1 square mile, occurs in the Copper Lakes region and at the head of

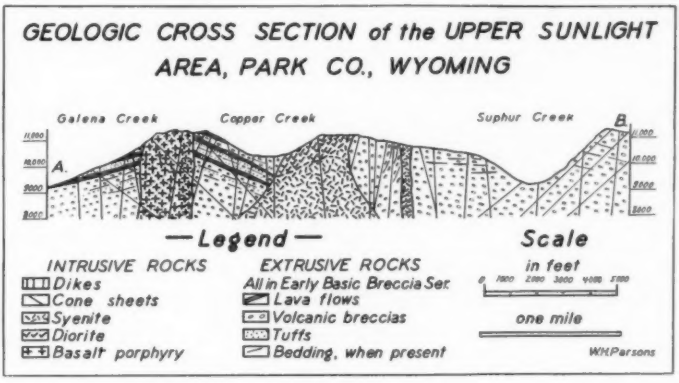

FIG. 8

Copper Creek (Fig. 7). Its contacts with the volcanic breccias are steeply flaring, although the southern arm of this stock is sill-like or tongue-shaped in structure. The stock was intruded late in the volcanic history of the region when the cones had reached their maximum heights and explosive activity was about over.

Just south of this stock, a coarse-textured syenitic to dioritic intrusive (1,000 feet wide) occurs, with a half-ring arcuate outcrop pattern about a mile across at the horns. This is considered to be a thick ring dike. The exposed volcanic breccias within the semicircle have steeply dipping to vertical bedding, suggesting that the area within the ring is a collapsed crater. An outcrop of monzogabbro in this crater area may be a small plug or pipe.

Numerous, narrow, andestic dikes cut the vent fillings, vent walls,

and to a lesser extent the stock in all directions, with no systematic arrangement. Vertical dikes radiate outward 5–7 miles from this center (see Fig. 4), the actual center being the middle of the stock. This radial system is far more perfect, extends farther in all directions, and contains more dikes than the White Mountain system. The actual number of dikes is in the thousands. The dikes average 4–6 feet in width, with a 20-foot maximum.

A few dikes, similar in composition and thickness to the dikes just described, strike approximately at right angles to the radial dike system and dip about 45° toward the volcanic center. These are cone sheets because of their concentric arrangement and inward dips. They are roughly centered about the dioritic (southwestern) part of the stock (see Fig. 7).

Associated with the various intrusives of the center are several small epithermal ore deposits[18] and considerable hydrothermal alteration of the country rock. The ores and alterations represent the last activity of the composite Sunlight volcano.

The flat-lying early basalt sheets outcrop 2 or 3 miles from this center to the east, south, and west at elevations averaging 11,400 feet. Within the area of the volcano, a half-dozen peaks, having no early basalt sheets, rise above 11,400 feet, one to almost 12,000 feet. These facts indicate that the Sunlight volcano stood as a high cone above the surrounding surface of the early basic breccia at the beginning of extrusion from fissures of the early basalt sheets, and was not covered by basalt.

The Sunlight volcano ejected most of the material of the early basic breccias within a radius of 2 or 3 miles and deposited some of the material in the breccia series even farther away.

Certainly, the Sunlight volcano was comparable in size to Vesuvius, though it is difficult to judge its exact size. Considering the diameter (4 or 5 miles) of the area that rose above the early basalt sheets, the fairly steep primary dips around some of the vents, and the coarse crystallization of the stock, it is safe to say that the pile must have risen 2,000–3,000 feet above the highest point today, or to well over 14,000 feet.

[18] Parsons, "The Ore Deposits . . . . ," *op. cit.*

MECHANICS OF INTRUSION

Cone sheets and radial dike systems owe their emplacement to an excess of magmatic pressure acting vertically upward on a relatively thin covering and to the doming of the roof of a magma reservoir. Richey and Thomas[19] have expressed this opinion. In the formation of cone sheets the upward force is applied in a restricted area; that is, the magma cupola has steep sides. Radial dikes form when the magma reservoir has a gently arched roof and a greater lateral than vertical extent. In the Sunlight volcano the radial dike system and cone sheets are centered about different parts of the small stock. The latter is part of a magma chamber of a shape best suited to the formation of radial cracks.

Anderson[20] writes that the tendency of a dome-shaped subterranean caldera to lift its roof forms tension fissures normal to the walls of the caldera. More than one set of surfaces, however, have this normal character. One set will radiate from the center as expressed in radial dike systems and the other will give rise to cone sheets.

Another kind of mechanics is also important in forming radial cracks in the Sunlight area, especially in connection with the White Mountain system. Iddings put it as follows:

> Radiating dikes may result from the tension produced in a circular wall of a conduit within the cone of a volcano by the hydrostatic pressure of the lava; . . . . The expansion of the walls by the pressure of the lava causes the walls to split in nearly vertical cracks that extend outward from the conduit.[21]

Around White Mountain the dikes cut the limestones near the actual vent, but beyond the vent a short distance the dikes occur only in the overlying volcanic breccias. The magma which formed these dikes must have moved horizontally outward from the conduit of the volcano in vertical cracks except within a half-mile of the vent where magma also ascended from deeper zones through the limestone.

[19] J. E. Richey and H. H. Thomas *et al.*, "The Geology of Ardnamurchan, Northwest Mull and Coll," *Mem. Geol. Surv. Scotland* (1930).

[20] E. M. Anderson, "Cone-Sheets and Ring-Dykes: The Dynamical Explanation," *Bull. Volcanologique*, Vol. I (2d ser., 1937), pp. 35–40.

[21] J. P. Iddings, *Problems of Volcanism* (New Haven: Yale University Press, 1914), p. 224.

The tension cracks near the Trail Creek vent are parallel to one another. Here there was no long-existing active volcano or an upthrusting body of magma. A single violent explosion through the crust formed a series of parallel fissures.

SUMMARY

One radial dike system was formed by horizontal injections of magma into vertical radial cracks, the latter being formed by violent explosions and upward pressure in the volcanic conduit. The other radial dike system was formed over a small stock where fissures developed normal to the stock's surface. A few of these normal fissures around one part of the stock took the form of cone sheets.

COMPARISON OF THE SUNLIGHT AREA WITH OTHER AREAS

BRITISH VOLCANIC REGION

In western Scotland and northeastern Ireland there are extensive areas of Tertiary volcanic rocks. Activity has been divided into four main episodes: (1) outpouring of great thicknesses of plateau basalts over a wide area from fissures and basaltic calderae, (2) establishment of central explosion vents and formation of vent agglomerates, (3) intrusion of plutonic and hypabyssal rocks at localized centers, and (4) intrusion of a great number of directionally oriented dikes. Excellent reviews of Tertiary igneous activity in Britain are given by Thomas in the Ardnamurchan memoir[22] and by Richey.[23]

Many points of similarity exist between this British region and the Absaroka volcanic field as studied in the Sunlight area. The formation of central vents and the localization of intrusives around these centers is characteristic of both areas. Local vent agglomerates, plugs, small stocks, laccoliths, and radial dike systems occur in Britain. However, ring dikes, cone sheets, and parallel dike swarms are the most characteristic features of the Scottish and Irish centers.

The predominance of radial dikes in the Absarokas rather than cone sheets and ring dikes as in Britain is probably due to the char-

[22] *Op. cit.*, pp. 50–64.

[23] J. E. Richey, "Some Features of Tertiary Volcanicity in Scotland and Ireland," *Bull. volcanologique*, Vol. I (2d ser., 1937), pp. 13–34.

acter of the rocks intruded as well as to the mechanics of intrusion. The British centers cut flat-lying plateau basalts, while the Absaroka centers cut volcanic breccias. The latter were probably poorly consolidated at the time of intrusion of the dikes and were piled in many cones comprising a more shatterable crust.

CRIPPLE CREEK VOLCANO

This denuded late Miocene volcano is a composite mass, formed mainly by explosive eruptions, of volcanic breccias and intrusive rocks (2 by 4 miles in extent) located in the Pike's Peak granite. Irregular pipelike intrusive bodies flare out at the top into small stocks. Late in the history of the volcano a local explosive eruption of basaltic breccia took place within the composite crater and formed a breccia pipe called the Cresson blowout.[24]

Several points of similarity exist between this volcano and the Sunlight volcano. The explosion breccias and basalt porphyry in the plugs at the head of Galena Creek are possibly similar in origin to the Cresson blowout. Dikes with a rudely radial arrangement occur at Cripple Creek. Metalliferous mineralization took place in both volcanoes, though weakly in the Sunlight area. At Sunlight the volcanic breccias of the cone are still partially preserved, making it difficult to determine the limits of a composite crater like that at Cripple Creek. However, the line marking the outer limit of hydrothermal alteration,[25] a roughly circular area (4 miles in diameter) including all vent structures and centralized intrusive bodies, might approximately coincide with the outline of such a composite volcano.

PETROLOGY

PETROGRAPHIC DESCRIPTION OF THE WHITE MOUNTAIN SERIES

A large part of the rocks making up the plugs and laccoliths of White Mountain are gabbros, olivine gabbros, or labradorite gabbro porphyries. These contain labradorite (Ab_{40} An_{60}), augite, biotite, magnetite, apatite, and varying amounts of olivine. All contain a little interstitial orthoclase. The labradorite gabbro porphyries contain large phenocrysts of basic labradorite or bytownite. All these rocks are dark colored and medium to coarse granitoid textured.

[24] G. F. Loughlin and A. H. Koschmann, "Geology and Ore Deposits of the Cripple Creek District, Colorado," *Colo. Sci. Soc. Proc.*, Vol. XIII (1935), pp. 217–435.

[25] Parsons, "The Ore Deposits . . . . ," *op. cit.*, p. 837 (map).

Parts of the plugs and laccoliths, particularly one of the laccoliths, consist of augite diorite and monzodiorite. These are fine-grained, light-colored rocks composed of andesine, oligoclase, orthoclase, augite, small amounts of biotite, magnetite, and apatite.

The dikes of the White Mountain suite have the following compositions: labradorite basalt porphyry, basalt porphyry, labradorite andesite porphyry, andesite porphyry, hornblende andesite porphyry, trachyandesite porphyry, and a few unusual absarokite porphyries, shoshonites, and leucite andesite porphyries. The basalts contain labradorite, augite, magnetite, apatite, and usually a little olivine. The andesites have labradorite, andesine, augite, magnetite, apatite, and a little interstitial orthoclase. The hornblende andesites have in addition black hornblende needles as phenocrysts. The trachyandesites carry oligoclase, orthoclase, augite, biotite, magnetite, and apatite. The absarokites[26] and shoshonites are orthoclase-bearing olivine basalts and the leucite andesites are orthoclase-rich andesites with a few leucite phenocrysts.

PETROGRAPHIC DESCRIPTION OF THE SUNLIGHT SERIES

The rock forming the plugs on Galena Creek is a basalt porphyry or glomeroporphyry (starlike clusters of labradorite phenocrysts) containing labradorite (Ab_{35} An_{65}), chloritized augite, magnetite, and a little serpentinized olivine.

The stock and ring dike are composed of medium granitoid-textured monzogabbro, diorite, monzodiorite, sodic syenite, and sodic quartz syenite. The monzogabbro contains labradorite, green hornblende, augite, and magnetite. The diorites and monzodiorites (one-half of the ring dike and one-third of the stock) carry andesine or oligoclase, orthoclase, augite, biotite, hornblende, magnetite, and apatite. The monzodiorites have only half as many mafic minerals as the diorites. The sodic syenites (two-thirds of the stock) carry albite, perthite, orthoclase, hornblende, biotite, magnetite, apatite, rare titanite, and some quartz (especially in the quartz syenites). Those portions which are porphyritic have phenocrysts of oligoclase (Ab_{80} An_{20}), and the mafic minerals occur in very small amounts.

The dikes and cone sheets of the Sunlight volcano have the follow-

[26] Iddings, "Geology of Yellowstone National Park," *U.S. Geol. Surv. Mon. 32*, Part II (1899), pp. 326–55.

ing compositions: basalt porphyry, andesite porphyry, andesite felsite, hornblende andesite porphyry, trachyandesite porphyry, and trachyte porphyry. The basalts have labradorite ($Ab_{35}An_{65}$), augite, rare olivine, magnetite, and apatite with some interstitial glass. The andesites contain labradorite ($Ab_{50}An_{50}$ to $Ab_{45}An_{55}$), augite, magnetite, and apatite. Rarely some black hornblende (green in section) occurs as phenocrysts. The trachyandesites consist of andesine or oligoclase, orthoclase, biotite, augite, hornblende, magnetite, and apatite. The trachytes are composed of oligoclase (as phenocrysts), albite, orthoclase, and rare biotite, magnetite, apatite, and quartz.

DISCUSSION OF CHEMICAL ANALYSES

Four chemical analyses (Nos. 1, 4, 5, and 6) were made in connection with this study. These together with ten others previously published[27] form the basis for the following discussion.

Analyses 1, 2, and 3 show that the rocks of the White Mountain suite are relatively basic ones characterized by a large amount of K_2O. Analysis 1, the typical olivine gabbro of White Mountain, proves to be an average olivine gabbro. Analysis 2, another facies of the same gabbro, is distinctly an orthoclase gabbro. Analysis 3 is identical with Iddings[28] absarokites. Sixteen miles northwest of the Sunlight volcano is located the Crandall Basin volcano on Hurricane Mesa.[29] The intrusives of this center are composed of orthoclase-rich gabbros and absarokites very similar in composition to the White Mountain rocks. The rocks of the two centers belong to the same petrographic province.

Eleven analyses (Nos. 4–14) of the Sunlight suite represent a fairly normal series of rocks, although some of the dikes are abnormally rich in orthoclase. Analysis 11 is similar to Iddings' shoshonites and Analysis 12, to his banakites.[30]

These two suites overlap each other in composition, the more basic rocks predominating in the White Mountain suite and the more acid in the Sunlight suite. The evidence is suggestive that both

[27] F. W. Clarke, "Analyses of Rocks from the Laboratory of the United States Geological Survey, 1880–1899," *U.S. Geol. Surv. Bull. 168* (1900), pp. 95, 97–98, and 109.

[28] "Geology of Yellowstone . . . . ," *op. cit.*, pp. 326–55.

[29] *Ibid.*, pp. 215–68.

[30] *Ibid.*, pp. 326–55.

TABLE 1

CHEMICAL ANALYSES OF ROCKS FROM THE SUNLIGHT AREA AND VICINITY

	1*	2	3	4	5	6	7	8	9	10†	11	12	13	14
SiO_2	46.95	53.57	47.32	52.28	55.98	57.78	63.07	64.40	66.64	47.17	51.17	52.47	57.73	63.24
Al_2O_3	18.53	17.78	11.22	15.29	16.62	17.93	17.47	16.90	16.22	17.85	16.14	18.23	18.93	17.98
Fe_2O_3	6.17	3.19	2.91	3.49	3.51	2.80	2.09	1.86	1.84	7.42	4.11	3.31	1.97	2.67
FeO	5.92	4.93	5.81	5.23	3.23	2.60	1.38	1.37	1.06	1.18	4.48	3.85	1.92	.85
MgO	4.96	4.36	15.96	6.57	3.41	1.87	1.44	1.13	1.25	6.54	4.82	2.85	.91	.63
CaO	10.51	6.22	7.11	6.58	4.91	3.65	2.27	2.60	2.41	10.12	7.72	4.56	2.78	.93
Na_2O	2.94	4.04	1.88	2.23	4.39	5.08	5.77	5.79	5.11	2.94	2.99	4.83	5.52	6.27
K_2O	1.23	3.04	3.79	2.64	4.23	4.48	4.59	4.56	3.86	.56	3.54	3.81	6.11	5.47
H_2O+	.68	.80	1.71	2.62	1.07	1.14	.43	.39	.55	.65	2.24	2.03	2.93	.80
H_2O-	.24	.27	.31	.58	.70	.06	.25	.16	.52	.65	.63	.68	.22	.37
CO_2	.....	.....	.13	.66	.14	.84	.....	.....	.....	.....	.....	1.01	.26	.....
TiO_2	.67	.89	.75	.78	.73	.64	.38	.23	.29	2.13	1.01	.97	.33	.38
P_2O_5	.64	.44	.61	.52	.58	.43	.18	.21	.16	.20	.48	.64	.25	.22
MnO	.14	.07	.11	.12	.14	.09	.03	.07	tr.	.....	.21	.15	.06	.04
SrO	.....	.13	.05	.....	.....	.....	.15	.14	.14	.....	.10	.11	.09	.03
BaO	.10	.21	.22	.....	.....	.22	.32	.27	.27	.....	.20	.23	.16	.25
V_2O_5	.....	.....	.....	.....	.....	.....	tr.	.....	.01	.....	.04	.03	.01	.01
FeS	.....	.....	.....	.....	.....	.....	.02	tr.?	tr.?	.....	.05	.04	.02	tr.
Li_2O	.....	tr.	tr.	.....	.....	.....	tr.	tr.	.....	.02	tr.	tr.	tr.	tr.
Cr_2O_3	.....	.....	tr.	.....	.....	.....	tr.	.....	tr.	.....	tr.	tr.	tr.	.....
NiO	.....	.....	tr.	.....	.....	.....	.....	.....	.....	.....	.01	tr.	tr.?	.....
Zr_2O_3	.....	.....	.....	.....	.....	.....	tr.	.02	.01	.....	.....	.02	tr.	tr.
Total	99.71	99.94	99.89	99.59	99.64	99.61	99.84	100.10	100.34	100.55	99.94	99.82	100.20	100.14

* No. 1: Also includes 0.03 per cent S.

† No. 10: Also includes 3.26 per cent Fe; 0.51 per cent SO_3.

1. Olivine gabbro, White Mountain stock, west side White Mountain. Analyst: A. Willman.
2. Gabbro, Sunlight Valley (Arnold Hague and T. A. Jaggar, Jr., *U.S. Geol. Surv. Bull. 168* [1900], p. 95). Analyst: H. N. Stokes.
3. Leucite absarokite, dike, Sunlight Valley (*ibid.*, p. 97). Analyst: H. N. Stokes.
4. Monzogabbro porphyry, ring dike, Sulphur Creek Basin. Analyst: A. Willman.
5. Monzodiorite, ring dike, Sulphur Creek Basin. Analyst: A. Willman.
6. Monzodiorite, transitional to syenite, ring dike, Sulphur Creek Basin. Analyst: A. Willman.
7. Syenite, stock, Copper Creek Basin (*ibid.*, p. 95). Analyst: W. F. Hillebrand.
8. Augite syenite porphyry, stock, Copper Creek Basin (*ibid.*). Analyst: W. F. Hillebrand.
9. Syenite porphyry, ring dike, Sulphur Creek Basin (*ibid.*). Analyst: W. F. Hillebrand.
10. Basalt porphyry, north spur Prospect Peak, Yellowstone National Park (*ibid.*, p. 109). Analyst: J. E. Whitfield.
11. Augite andesite, Dike Mountain (*ibid.*, p. 98). Analyst: W. F. Hillebrand.
12. Trachyandesite, Dike Mountain (*ibid.*). Analyst: W. F. Hillebrand.
13. Biotite trachyte, Dike Mountain (*ibid.*). Analyst: W. F. Hillebrand.
14. Trachyte, Dike Mountain (*ibid.*). Analyst: W. F. Hillebrand.

TABLE 2

NORMS OF ROCKS FROM THE SUNLIGHT AREA AND VICINITY*

	1†	2	3	4	5	6	7	8	9	10‡	11	12	13	14
Quartz	.....	.....	.....	6.33	.90	3.07	5.70	7.20	15.96	.....	.....	.....	.....	3.12
Orthoclase	7.23	17.79	22.24	15.57	25.02	26.69	27.24	27.24	22.80	3.3	20.57	22.2	36.14	32.80
Albite	24.63	33.54	9.43	18.86	37.20	42.97	48.73	48.73	42.97	24.6	25.15	39.8	38.77	52.92
Anorthite	33.64	21.68	11.12	23.91	13.07	10.29	8.06	6.67	10.01	33.9	20.29	17.2	8.62	3.61
Nepheline	.....	.....	3.41	.....	.....	.....	.....	.....	.....	.....	.....	.3	4.26	.....
Corundum	.....	.....	.....	.....	.....	.82	.....	.....	.....	.....	.....	.....	.....	.31
Diopside	11.61	4.94	16.27	.68	5.54	.....	1.94	4.82	.86	10.4	11.98	1.2	2.72	.....
Hypersthene	4.06	12.49	.....	21.61	8.02	6.17	2.83	1.16	2.70	11.8	8.41	.....	.....	1.60
Olivine	5.41	.31	28.07	.....	.....	.....	.....	.....	.....	.....	1.18	7.5	1.76	.....
Magnetite	9.05	4.64	4.18	5.10	5.10	4.18	3.02	2.78	2.55	.....	5.80	4.6	3.02	1.62
Ilmenite	1.37	1.67	1.37	1.52	1.37	1.22	.76	.40	.61	2.3	1.82	1.8	.61	.76
Hematite	.....	.....	.....	.....	.....	.....	.....	.....	.16	7.4	.....	.....	.....	1.60
Apatite	1.68	1.01	1.34	1.34	1.34	1.01	.34	.34	.34	.....	1.01	1.4	.67	.67
Calcite	.....	.....	.....	1.50	.30	1.90	.....	.....	.....	.....	.....	.....	.....	.....
Total (*minus* H_2O)	98.80	98.07	96.43	96.42	97.86	98.32	98.62	99.34	98.96	95.7	96.21	96.0	96.57	99.01

* Names and locations of specimens as in Table 1. † No. 1 also includes 0.12 pyrite. ‡ No. 10 also includes 2.0 titanite.

suites differentiated from a common parent magma and are members of the same petrographic province.

ORIGIN AND DIFFERENTIATION OF THE ROCK TYPES

The general course of crystallization of the magma of the White Mountain and Sunlight rocks was a normal one following the basalt-trachyte line of descent. It is theoretically possible in dry melts approaching the composition of basaltic liquids to produce, under different conditions of cooling, an oversilicated or granitic residuum, or a neutral or undersilicated one, the syenitic residuum. Bowen[31] suggests that a different rate of cooling will influence the composition of the residual liquid. Barth[32] states that a very slight variation in the composition of the initial liquid may, by fractional crystallization, give rise sometimes to an oversilicated and sometimes to an undersilicated residuum.

Bowen[33] has suggested that biotite forming early and sinking in the magma chamber would be resorbed at lower and hotter levels, thereby forming a liquid rich in femic molecules and potash. This might give rise to the absarokites which have an excess amount of olivine and augite as phenocrysts and an excess of potash as interstitial leucite or orthoclase. Partially resorbed biotite is seen in many of the absarokites and shoshonites. However, this theory fails to explain the banakites, the other end member of the potash-rich rock series, which contain labradorite with leucite or orthoclase but very few mafic constituents.

Another hypothesis for the explanation of some of the orthoclase-rich rocks based on the principle of fractional crystallization is suggested as follows. Some unusual conditions obtained during crystallization which permitted the early separation of leucite crystals along with numerous labradorite crystals. These crystals rose and were separated from the parent magma before they could react with it. Most of the leucite was then resorbed, giving rise to a potash-rich interstitial liquid which later recrystallized as tiny groundmass

[31] N. L. Bowen, *The Evolution of the Igneous Rocks* (Princeton: Princeton University Press, 1928), pp. 236–40.

[32] Tom F. W. Barth, "The Crystallization Process of Basalts," *Amer. Jour. Sci.*, Vol. XXXI (1936), pp. 321–51.

[33] *Op. cit.*, pp. 271–72.

leucite crystals or more commonly as interstitial orthoclase, depending on the percentage of residual silica. Leucite phenocrysts, which are occasionally seen, are always much corroded.

SUMMARY OF VOLCANIC HISTORY

We may list the following episodes in the volcanic history of the Sunlight area: (1) initiation of pyroclastic volcanic activity from a great number of explosive vents, building up a thick series of volcanic breccias; (2) the concentration of activity into a few main volcanoes and the building of cones over these centers; (3) the outpouring of basalt sheets as fissure eruptions; (4) the injection of plutonic bodies in the bases of the volcanoes; and (5) the injection of thousands of radial dikes outward from the extrusive and intrusive centers.

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THE BELL RIVER COMPLEX, NORTHWESTERN QUEBEC

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ABSTRACT

A complex of basic rocks in the Waswanipi District of northwestern Quebec is described, and the interpretation is advanced that it is a deformed lopolith of the same general character as the Stillwater complex of Montana and the Bushveld complex of the Transvaal. The rocks of the complex are severely altered, and the alterations—serpentinization, hornblendization, and chloritization—are described. The serpentinization is thought to have been caused by deuteric processes within the basic mass, and the other changes are believed to have been brought about by hydrothermal solutions derived from intrusive quartz-diorite.

INTRODUCTION

The Bell River complex is in the Waswanipi region of northwestern Quebec. The general geology of the surrounding district has been described in an article by the writer[1] and need not be treated here. The distribution and relative ages of the rocks are shown in Figure 1, in which the complex is represented as anorthosite in the legend. The anorthosite intrudes Keewatin-type volcanic complex and is intruded by batholithic masses of quartz-diorite. Fresh-appearing lamprophyre and diabase dikes cut all the other types but were not seen in contact with each other.

The occurrence of a large amount of basic rock outcropping for 20 miles along the Bell River was noted by Bancroft in 1912,[2] who called the rock a gabbro-diorite and concluded that it was originally an anorthosite and was related to the Chibougamau anorthosite.[3] In 1931 Lang correlated the rock with the Chibougamau mass and called it an oligoclase anorthosite,[4] following the usage of Mawdsley

[1] B. C. Freeman, "Replacement Shells around Batholiths in the Waswanipi District, Northwestern Quebec," *Jour. Geol.*, Vol. XLVI (1938), pp. 681–99.

[2] J. Austin Bancroft, "A Report on the Geology and Natural Resources of Certain Parts of the Drainage Basins of the Harricanaw and Nottaway Rivers, to the North of the National Transcontinental Railway in Northwestern Quebec," *Quebec Bur. Mines Rept. on Mining Operations* (1912), pp. 131–98.

[3] *Ibid.*, p. 155.

[4] A. H. Lang, "Waswanipi Lake Area, Quebec," *Geol. Surv. Canada Summ. Rept.* (1932), Part D, p. 40.

for the Chibougamau occurrence.[5] All investigators have noted that it varies in texture and mineralogy from place to place and that it has been altered, but it has never been described at all fully.

Three areas underlain by anorthosite and related rocks are indicated in Figure 1. These three masses are separated by intrusive quartz-diorite, and numerous xenoliths too small to be shown on the map occur within quartz-diorite. Undoubtedly there was once a single mass of anorthosite and related rocks about 40 miles long and 12 miles wide. Outcrops are so few that there may be an error of as much as ½ mile in the location of the boundaries in many places, but the size and shape of the complex must be about as shown. The structure of the Keewatin rocks could be determined in only a few outcrops, but, where determined, the strike of the schistosity and of the lava flows parallels the long axis of the complex. Banding within the complex, interpreted as primary, also parallels the long axis of the mass except near the ends where it parallels the contact. Probably the intrusion was concordant. Dips, wherever observed, both in the rocks of the complex and in the country rock are vertical or nearly so. The map, therefore, represents a cross section through the mass.

DESCRIPTION OF THE COMPLEX

Of the three large masses of basic rock, the largest, the Bell River anorthosite proper, is about 25 by 10 miles in size and extends across the Bell and Olga rivers. The next largest is 10 by 6 miles in size and lies south of Gull Lake. The Dalhousie Mountains are composed of the Gull Lake anorthosite which has been very severely altered and is riddled by dikes and irregular masses of quartz-diorite. The smallest mass is about 2 by 1 miles in size and is separated from the Bell River mass by about a mile of quartz-diorite. It is apparently a xenolith and probably is irregular, not smoothly oval in outline as shown on the map. The rock in one outcrop near the center of this mass is comparatively unaltered and is a bytownite (An_{80}) anorthosite (see Fig. 5, *D*).

The Bell-Olga mass retains some of the original differences in texture, structure, and composition that probably characterized the

[5] J. B. Mawdsley, "Lake David Area, Chibougamau District, Quebec," *Geol. Surv. Canada Summ. Rept.* (1927), Part C, p. 8.

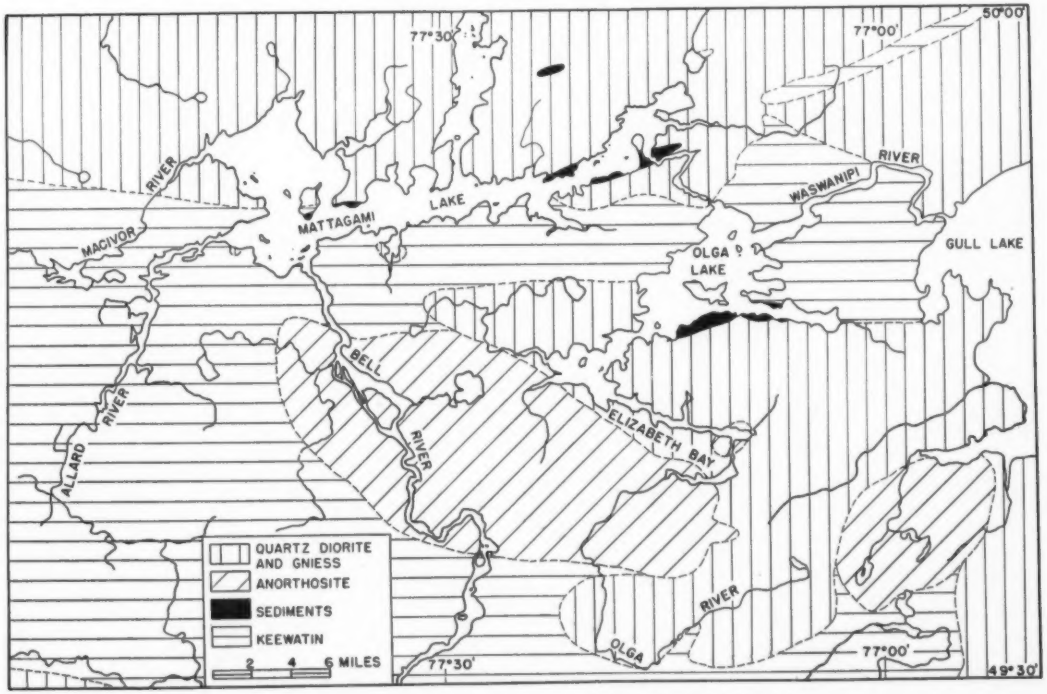

FIG. 1.—Map of the northwestern part of the Waswanipi district

whole mass. The Gull Lake mass has been so altered by the addition of new material, mostly hornblende, that original differences are

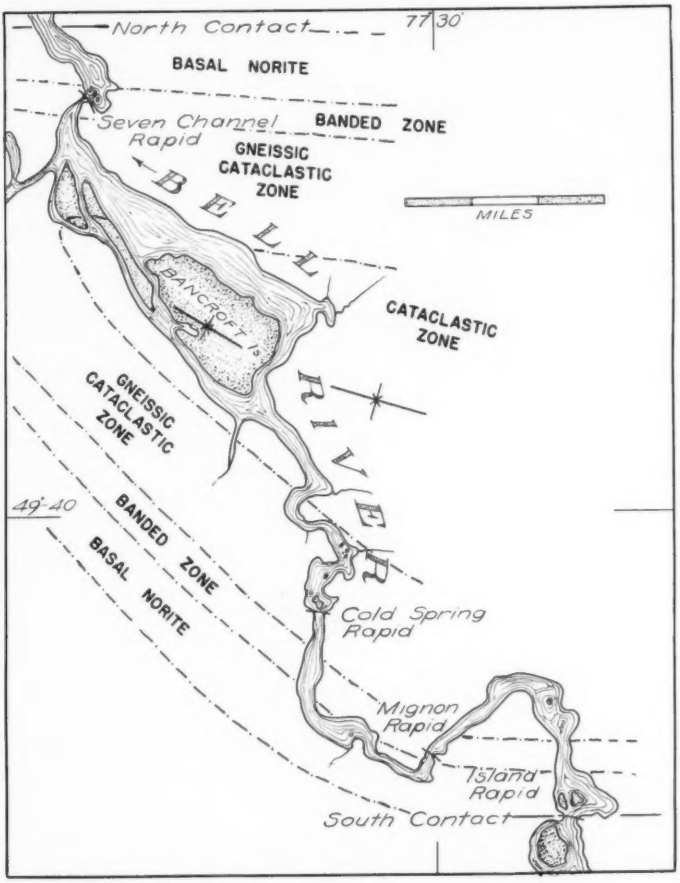

FIG. 2.—The course of the Bell River across the complex

obscured. It represents an end product seen only sparingly in the Bell-Olga mass. Therefore, a section across the Bell-Olga mass along the Bell River gives the most significant information about the original composition of the mass.

This section along the Bell River comprises four major units, as follows: (1) a basal norite, 4,000 feet thick; (2) a banded zone, 2,500 feet thick; (3) a gneissic cataclastic zone, 8,500 feet thick; and (4) a cataclastic zone, not particularly banded, at least 5,000 feet thick. In Figure 2 the course of the Bell River across the complex is shown, and the distribution of the four units on either side of a central

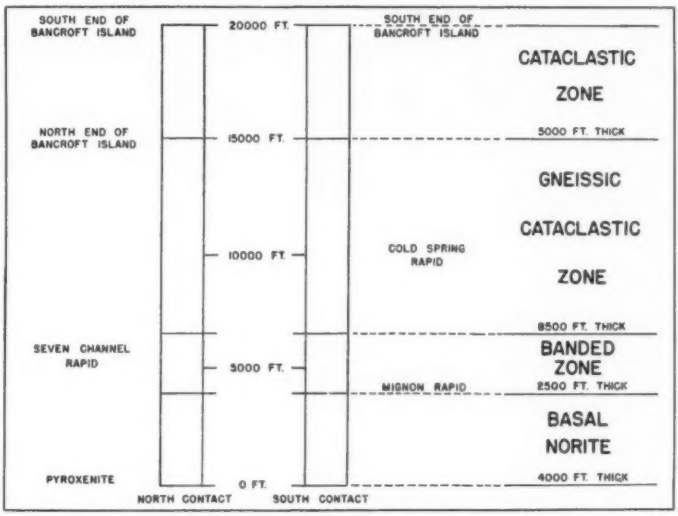

FIG. 3.—Zones in the Bell River complex

synclinal axis is indicated. The structure is believed to be a tight syncline with nearly vertical limbs. In Figure 3 the zones are arranged in two columns, showing the thickness of each zone and the relation of the zones to the contact and to some geographic features. A detailed description of the outcrops along the river from the south contact, at the Island rapid, to the north contact follows, and from variations in petrography, texture, and structure the presence of the four zones is indicated.

At the south contact there are inclusions of altered greenish-colored, fine-grained rock in norite. These inclusions may be fragments of intruded lava complex or fractured pieces of a basal chill

zone formed by folding of a lopolith while the main part of the mass was not yet crystallized.

At the south side, for about 2,500 feet across the edge of the mass, most of the rock is a coarse norite, with a faint diabasic texture. The grain size is variable from place to place, the individuals ranging in length from about 3 to 15 mm. The essential primary constituents were plagioclase and pyroxene in laths and irregular clots of embayed laths up to 2 inches across. The plagioclase is white and chalky to slightly greenish, though there are a few glassy crystals with good twinning lines. The pyroxene is dull black to the yellow-green characteristic of serpentine. A little titaniferous magnetite is interstitial to the other components, principally pyroxene, and a few specks of pyrite are scattered throughout the rock (see Fig. 4, *B*).

Some of the rock is poor in ferromagnesian constituents and approaches anorthosite in composition. These anorthosite masses are irregular in shape, are up to several feet long, and have indefinite, gradational boundaries against the norite.

Some parts, the boundaries of which were not seen, are very similar in appearance to the quartz-diorite intrusive into the complex farther east. These are oligoclase-rich rocks with a very little hornblende, 1 per cent or so, and about 15 per cent quartz, most of which shows replacement boundaries against oligoclase. The oligoclase is mostly untwinned and badly altered.

White, pure quartz veins are very numerous but small; the largest one seen was only 4 inches wide.

The norite in thin section is seen to have been composed of labradorite ($Ab_{40}An_{60}$) and hypersthene. The labradorite is altered to a mat of secondary minerals. Quartz, zoisite, and calcite, in grains and irregular masses up to 2 mm. across, occur in a base of clear untwinned albite. Ghosts of the original twinning are present in some crystals, and the habit and twinning angles agree with those of median labradorite. Such palimpsests of the original plagioclase twinning are rather common throughout the complex. The index is usually much lower than that of labradorite and probably indicates the correct present composition of the plagioclase (see Fig. 5, *C*). Calcite is very prominently developed in the plagioclase, so much so that most specimens effervesce freely when hydrochloric acid is

placed on them. The lime may have been derived from the basic plagioclase when it was altered.

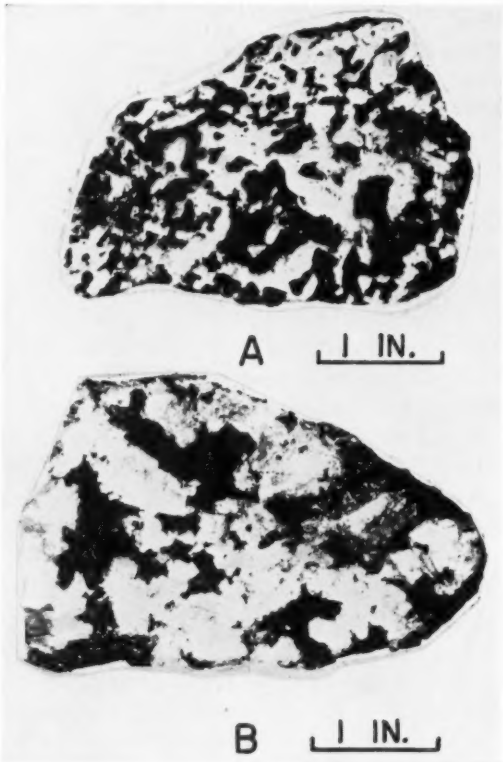

FIG. 4.—Textures in norite. *A*, norite from the banded zone; *B*, norite from the basal norite. White and light-gray plagioclase and dark-gray to black altered hypersthene. Some magnetite is present associated with the hypersthene but is not distinguishable in the photographs from the pyroxene.

Remnants of unaltered hypersthene are left in the centers of some individuals, but almost all of it is altered to antigorite, bastite, or some other serpentine-like material. The hypersthene contains abundant specks of magnetite in lines along cleavage planes and has

the schiller characteristic of orthorhombic pyroxenes. In the alteration most of these inclusions become opaque, white masses of leucoxene, and the schiller effect is not marked in antigorite. Calcite occurs in minute flakes in the antigorite and in hypersthene, near the outside edge of remnants. Tiny flakes of hornblende also occur in the antigorite (see Fig. 5, *F*).

The opaque minerals were studied in polished specimens. Some magnetite occurs in rounded grains and irregular masses showing a pronounced development of the octahedral parting. Other masses are darker in color, lack the octahedral partings, and contain exsolution rods and laths of ilmenite. These grains are probably rich in TiO_2 (see Fig. 6, *A*). The magnetite is replaced along the octahedral partings and the more massive titaniferous magnetite is replaced peripherally by hematite and a still darker gray mineral which is magnetic, anisotropic, and shows no internal reflection (see Fig. 6, *B*). It very probably is maghemite. This name was suggested by Wagner[6] for magnetic ferric oxide, i.e., magnetite which had been oxidized to the composition of hematite without losing the magnetism or acquiring a red streak. Walker analyzed maghemite from the Bushveld complex and concluded that it had the composition $(Fe, Ti)_2O_3$.[7]

Specks of titaniferous magnetite about 0.1 mm. in size occur abundantly throughout the pyroxene but are nearly all replaced by maghemite so that the original mineral is left only as fragments scattered throughout the maghemite.

Pyrite in 0.1 mm. specks is distributed throughout all the other constituents, but is mainly in plagioclase. It also occurs as grains up to 1 mm. across. Every one seen was surrounded by a rim of maghemite or hematite. In the larger grains, evidence of replacement is good and peripheral replacement of the pyrite specks is assumed. The larger grains of pyrite contain angular pieces of chalcopyrite. One grain 1 mm. across contained six chalcopyrite fragments. These relationships are shown in Figure 6, *D*.

[6] Percy A. Wagner, "Changes in the Oxidation of Iron in Magnetite," *Econ. Geol.*, Vol. XXII (1927), pp. 845–46.

[7] T. L. Walker, "Lodestone from Bon Accord, Transvaal," *Univ. Toronto Studies Geol. Series, No. 29* (1930), pp. 17–19.

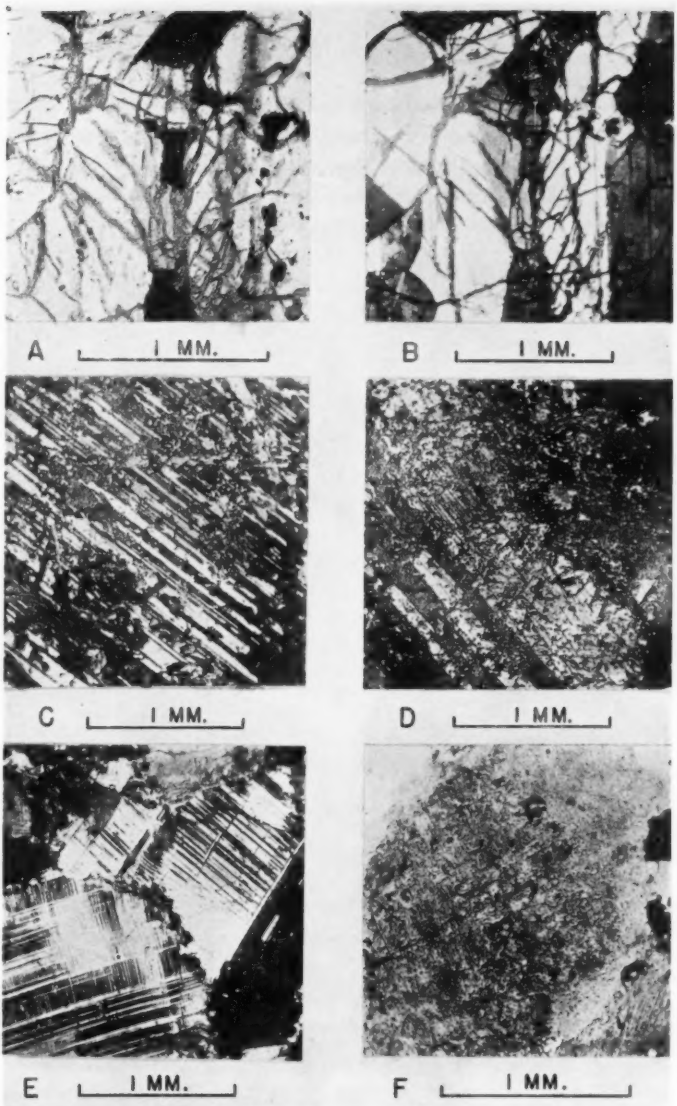

FIG. 5.—*A*, Ordinary light. Cataclastic labradorite-bytownite (An_{70}) cut by chlorite. From the Seven Channel Rapid. *B*, Nicols crossed; same field as *A*. *C*, Nicols crossed. Plagioclase with the habit and extinction angle of labradorite (An_{60}) but with the index of oligoclase-albite (An_{10}). The alteration is characteristic of much of the relatively fresh plagioclase. From the south end of Bancroft Island. *D*, Nicols crossed. Bytownite from the xenolith on the south arm of Olga Lake. Comparatively unaltered. *E*, Nicols crossed. Cataclastic labradorite showing both albite and pericline twinning. From the south bay of Gull Lake. This rock is extensively replaced by hornblende. *F*, Ordinary light. Hypersthene remnant in antigorite. Dark-gray spots are hornblende flakes. From the basal norite near the Island rapid.

Magnetite has also been replaced along the octahedral partings by nonopaque secondary minerals, calcite, zoisite, hornblende, quartz. Textural relationships between the iron-bearing minerals are shown in Figure 6, *A* and *B*.

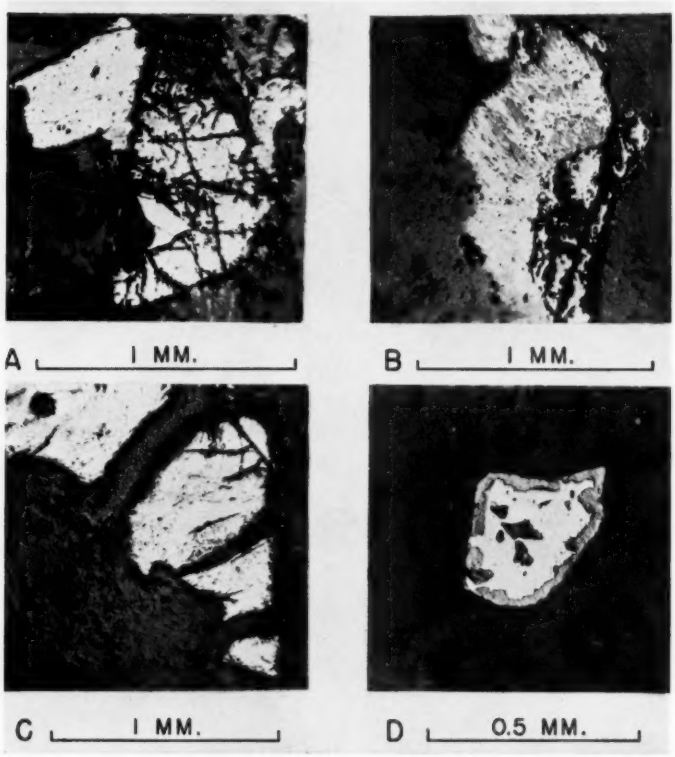

FIG. 6.—*A*, Magnetite (white) cut along octahedral partings by nonmetallics (dark gray) which have torn out in polishing leaving holes (black). Titaniferous magnetite is light gray and massive. *B*, Magnetite (light gray) more extensively replaced along octahedral partings than in *A*. Titaniferous magnetite replaced irregularly by maghemite (dark gray). Nonmetallics are darkest gray. *C*, Magnetite cut by nonmetallics, principally chlorite. *D*, Pyrite (white) enclosing angular grains of chalcopyrite (dark gray to black) and replaced by a rim of maghemite.

For 4 miles downstream, around a pronounced bend in the river, there are a few outcrops, all of altered norite and/or gabbro. At the Mignon Rapids there are excellent outcrops of phenomenally banded rock which is about 4,000 feet from the outer edge of the complex. At least twenty bands from 8 inches to 2 feet wide are present and most of the bands are persistent. Some, however, split and others coalesce. The strike of the banding is N. 60° W., or about parallel to the contact, and the dip is vertical.

The most abundant type is a norite composed of the same minerals in about the same amounts as the norite described above, but varying in grain size from about 1 to 2 mm. at the southwest edge to very coarse, about 2 cm., at the northeast side. The texture of a medium-grained part of this norite is shown in Figure 4, *A*. The maximum width of these norite bands is about 3 feet. Northeast of the coarse norite there is a band of altered, wedge-shaped pyroxene crystals, probably originally hypersthene or bronzite, with small amounts of plagioclase. The pyroxene crystals are 4–6 inches long and lie with their long axes parallel to the norite band, but are piled without any definite orientation within this plane. The plagioclase is in crystals about 1 cm. long. Next to the pyroxenite is a band of $\frac{1}{2}$ mm. grained anorthosite—almost pure plagioclase. The pyroxenite and anorthosite are each about 4 inches thick and are quite regular. The thickness of the norite bands does not vary much in any one band, but different bands are markedly different in width, some being only a few inches thick.

This remarkable alternation of bands of similar composition and texture is regarded as a primary structure. The anorthosite and pyroxenite bands considered together have approximately the composition of the norite part. The differences are thought to be due to nearly complete segregation of the principal constituents in a process of crystallization differentiation and crystal settling. The hypothesis favored is the same as that held by many geologists for the origin of the regional banding in the differentiated or critical zone of the Bushveld norite.[8]

[8] A. L. Hall, "The Bushveld Igneous Complex of the Central Transvaal," *Geol. Surv. South Africa Mem. 28* (1932), pp. 279–81.

Below the Mignon Rapids the river swings closer to the contact and parallels it about 2,500 feet away. The rock is again a norite.

The course of the river again changes, this time to north, but there are no outcrops for a mile and a half. This outcrop-free section is followed by a 2-mile stretch with numerous exposures. At the south side, at Cold Spring Rapids, the rock is gneissic norite, striking east, with grains 2–4 mm. in size. The banding is caused by different proportions of pyroxene and plagioclase, bands of light norite over half plagioclase alternating with dark bands containing less than 25 per cent plagioclase. The bands vary in width from an inch to a foot. Specks of pyrite are abundant throughout this type, which is about 10,000 feet from the contact.

From Cold Spring Rapids for 2 miles downstream and 15,000 feet from the contact, the rock is norite to anorthositic norite and much of it is gneissic. The anorthositic norite contains less than 20 per cent dark minerals, now mostly light-green chlorite with quartz and calcite, but an occasional individual of antigorite remains. The feldspar is typically cataclastic and almost completely altered.

Below this stretch is another 2-mile-long outcrop-free section of the river, the next outcrops being on Bancroft Island. The rock near the south end of the island is about 20,000 feet from the edge of the complex and is about in the center of the intrusive mass. It is typically a cataclastic anorthositic norite with labradorite as the plagioclase. Figure 5, *C*, is a photograph of plagioclase from Bancroft Island that is much fresher than most plagioclase found there. Most of the rock is massive, but some is gneissic, formed of pyroxene-rich and pyroxene-poor bands 2–6 inches in width, striking N. 60° W. There are also pegmatitic bands formed of a central zone of plagioclase in crystals up to 1 foot in length bordered by radially arranged, platy, serpentinized pyroxene crystals up to 6 inches in length. These crystals are wedge shaped, up to 2 inches in thickness at one end, and a number of them taper in toward common centers.

At the north end of the island, about 15,000 feet from the north side of the complex, most of the rock is again cataclastic, anorthositic norite, but there is also some pyroxenite formed of crystals of altered pyroxene 4–6 inches long. The masses of nearly pure pyroxene have very indefinite and irregular boundaries.

The next significant group of outcrops occurs at the sharp bend 3 miles below Bancroft Island, where the river falls over the Seven Channel Rapids. At the south side of the rapids the rock is about 6,000 feet from the edge of the complex and forms a series of alternating bands of severely altered and less altered anorthositic norite. The members are 6–12 feet wide, are vertical, and strike N. 80° E. Also some of the norite is banded like that at the Mignon Rapid, 4,000 feet from the south side. The fine-grained member is at the north, and the grain size increases toward the south. In both these occurrences the grain size increases toward the center of the mass and decreases toward the outer edge, which is interpreted as being originally at the bottom of the intruded mass of magma. The rock here is much more severely altered than that at the Mignon Rapid, and the alteration is of a different type. Not much serpentine was formed but rather considerable amounts of chlorite, which has replaced the altered pyroxenes and also the anorthosite along cracks caused by cataclastic action (see Fig. 5, *A* and *B*). The separation of magnetite grains into segments by chlorite replacement is shown in Figure 6, *C*.

On the north side of the rapids, about 5,000 feet from the north side of the complex, the rock contains 5–10 per cent of titaniferous magnetite and is very markedly chloritized. The rock is variable in grain size and has a parallel texture formed of laths of labradorite-bytownite, about An_{30}, and altered pyroxene. The plagioclase laths are cracked and replaced by chlorite seams but are quite fresh. There are, however, spots of alteration products, largely calcite or other carbonate with some zoisite, within labradorite. The pyroxene is almost all altered. A few irregular areas of original diallage remain, but practically all the mafic material forms mats of chlorite with numerous specks of calcite and a little zoisite and epidote. There are also some small flakes of hornblende—some brown and some blue-green—in the chlorite. Chlorite replaces titaniferous magnetite as well as the other constituents (see Fig. 6, *C*).

Outcrops are poor between this point and the contact which is exposed on the east bank of the river. The country rock is a dacite of the Keewatin-type volcanic complex, intruded by a severely altered pyroxenite and some norite.

The pyroxenite is a dark-greenish-gray rock with abundant glistening black crystals of hornblende up to $\frac{1}{8}$ inch, and irregular rounded areas of a yellowish-green aggregate up to $\frac{1}{2}$ inch, the whole in a fine-grained ($\frac{1}{2}$ mm.) base. In thin section the yellowish spots are seen to be aggregates of calcite and epidote, with zoisite, chlorite, quartz, and acid plagioclase in angular or ragged individuals 0.2–0.5 mm. across. The hornblende is twisted, poikilitic, and partly chloritized. The chadacrysts in the poikilitic hornblende are of the same material as the spots, but in different proportions—epidote being most abundant and calcite the next largest in amount. Possible palimpsests of original pyroxene are present as aggregates of epidote, calcite, chloritized uralite, and quartz. Figure 7, *F*, shows one such occurrence. The base is formed of the same materials as the yellowish spots. The texture and mineral composition are quite similar to those of the altered pyroxenes in the exposures at the Seven Channel Rapid 4,000 feet away, except for the presence of the quartz and acid plagioclase. These two constituents are probably due to the alteration of primary basic plagioclase in the original rock or have been introduced hydrothermally. I believe that the rock represents a mafic segregate at the bottom of a concordant intrusion.

The similarity of this complex to the basic rocks in the Chibougamau district was noted above. Recent work by Norman has shown the occurrence of a remarkable series of ultrabasic to dioritic sills north of the Chibougamau anorthosite and gabbro in the Opemisca district about 100 miles east of this region.[9] These sills are thought to be genetically related and range from ultrabasic at the base of the series to siliceous at the top. The base of each sill is also more basic than the top of that sill. The sills are believed to have been injected into nearly flat-lying Keewatin rocks and the whole was later folded into a syncline.[10]

The description of the rock types, and the zoning as indicated in Figures 2 and 3, bring out a remarkable similarity between this mass

[9] G. W. H. Norman, "Preliminary Report, East Half Opemisca Map-Area, Quebec," *Geol. Surv. Canada Paper 31–11* (April, 1938); "Preliminary Report, West Half Opemisca Map-Area, Abitibi Territory, Quebec," *Geol. Surv. Canada Paper 38–11* (April, 1938).

[10] Norman, "Preliminary Report, East Half Opemisca Map-Area," *op. cit.*, p. 6.

and the Bushveld[11] and Stillwater[12] complexes, both of which are regarded as lopoliths. The Bell River complex, because of this marked resemblance to these two lopoliths, is interpreted as a deformed lopolith. If this conclusion is correct, there is a possibility of the occurrence of important deposits of chromite, platinum, and nickel ores in this complex.

METAMORPHISM OF THE COMPLEX

Two grades of metamorphism are observed in the rocks of the complex. In the western part the mineral alteration, while severe, did not obliterate the original structures and textures of the rock. In the eastern part the changes were so great that a decidedly different rock type was produced.

CHANGES IN THE WEST PART OF THE COMPLEX

These changes were noted in the foregoing description of the outcrops along the Bell River. They fall into two groups that may be called, for convenience, serpentinization and chloritization.

1. *Serpentinization.*—Except near the Seven Channel Rapid, almost all the original pyroxene of the complex has been altered to serpentine with subordinate amounts of hornblende and calcite. Titaniferous magnetite is a common accessory mineral, associated more closely with pyroxene than with plagioclase. In the serpentinized areas there are numerous white, opaque masses resembling leucoxene. Probably leucoxene has formed from the titaniferous magnetite, the iron of the magnetite going into the hornblende as it was formed. The plagioclase is altered, in part or entirely, to a mat of epidote, zoisite, calcite, quartz, and acid plagioclase. In some places there are palimpsests of basic plagioclase in acid plagioclase having the index of oligoclase-albite. The alteration is to materials usually constituting saussurite, but the alteration products are in very much larger particles than in typical saussuritization (see Fig. 5, *C*).

[11] A. H. Hall, "The Bushveld Igneous Complex of the Central Transvaal," *Geol. Surv. Union of South Africa Mem. 28* (1932), pp. 271, 291, and 301.

[12] J. W. Peoples, "Gravity Stratification as a Criterion in the Interpretation of the Structure of the Stillwater Complex, Montana," *Rept. XVI Int. Geol. Congress* (1933), p. 358.

2. *Chloritization.*—Near the Seven Channel Rapid, close to the north side of the complex, the alteration has produced a black rock that resembles fresh diabase. The plagioclase is cataclastic but quite fresh except for extensive replacement by chlorite along the cracks (see Fig. 5, *A* and *B*). The pyroxene is altered to aggregates of hornblende, chlorite, calcite, and quartz. The black color of the rock is due to the presence of large amounts of chlorite, which has replaced the earlier-formed hornblende and the labradorite. Titaniferous magnetite is an important primary constituent and is also extensively replaced by chlorite (see Fig. 6, *C*).

CAUSE OF THE ALTERATION

These alterations resemble those described by Hess in his study of the alterations of the basic and ultrabasic rocks of the Appalachian region.[13] He concludes that serpentine is formed by deuteric or hypohydrous processes and that its formation preceded hydrothermal action.[14]

The formation of hornblende, actinolite, chlorite, talc, and carbonate is characteristic of the process called steatitization by Hess.[15] The chloritization of the rock near the Seven Channel Rapid resembles steatitization, except for the lack of talc, and the chloritization of this rock appears to have been caused by the same processes as steatitization. Hess concludes that these processes are hydrothermal, later than the hypohydrous processes, and that the alteration represents a second-cycle change due to solutions derived from an acid magma.[16]

The alteration of the plagioclase probably occurred along with the serpentinization of the pyroxene since the labradorite of the chloritized rock is fresh. It may of course have been due to different processes from those causing the pyroxene alteration. The large amount of carbonate in the alteration products indicates the presence of abundant carbon dioxide, which is typical of hydrothermal

[13] H. H. Hess, "Hydrothermal Metamorphism of an Ultrabasic Intrusive at Schuyler, Virginia," *Amer. Jour. Sci.*, Vol. XXVI (1933), pp. 378–408; "The Problem of Serpentinization and the Origin of Certain Chrysotile Asbestos, Talc, and Soapstone Deposits," *Econ. Geol.*, Vol. XXVIII (1933), pp. 634–57.

[14] "The Problem of Serpentinization . . . . ," *op. cit.*, p. 652.

[15] *Ibid.*, p. 635.

[16] *Ibid.*, p. 655.

solutions. If hydrothermal solutions affected the feldspar they must have been different from those causing chloritization, since the labradorite of the chloritized rock is so little changed.

CHANGES IN THE EASTERN PART OF THE COMPLEX

The rock south of Gull Lake is markedly different in appearance from that along the Bell River. It is formed principally of acid labradorite (An_{50}) and fresh hornblende, black and glistening in hand specimen and strongly pleochroic in blue-green and yellow-green colors in thin section. The labradorite is in strained and crushed aggregates showing mortar texture (see Fig. 5, *E*). Hornblende, in needles and small rounded masses, occurs abundantly within the labradorite, as shown in Figure 7, *C* and *D*. Small quartz blebs are present along minute cracks in labradorite resembling beads on a string, and quartz also occurs as larger grains in the crushed zones of the labradorite. In Figure 7, *E*, some quartz blebs and hornblende needles are shown in labradorite. The labradorite is crowded with tiny fragments of these materials. They replace the labradorite and are interpreted as a hydrothermal product.

Most of the hornblende occurs as large, irregular, poikilitic individuals with quartz chadacrysts (see Fig. 7, *A* and *B*). The large hornblende grains have replacement relationships against labradorite.

Magnetite occurs as blebs and rounded, irregular grains, mostly in hornblende but also in labradorite. It has relationships similar to those of quartz and is regarded as having been formed at the same time and in the same manner as quartz.

Hornblende, quartz, and magnetite definitely replace plagioclase and were introduced into the rock after the plagioclase was solid. Any original pyroxene present has disappeared in the general alteration and replacement of the rock. The age relationships of quartz and hornblende are in doubt. The quartz chadacrysts in hornblende may be older, the same age as, or younger than, the hornblende. Blebs within poikiloblasts are generally regarded as older than the host and have resisted change while the host mineral was developing. This is not necessarily so, because quartz replacing epidote in some phases of the quartz-diorite of the region assumes a chadacrystic

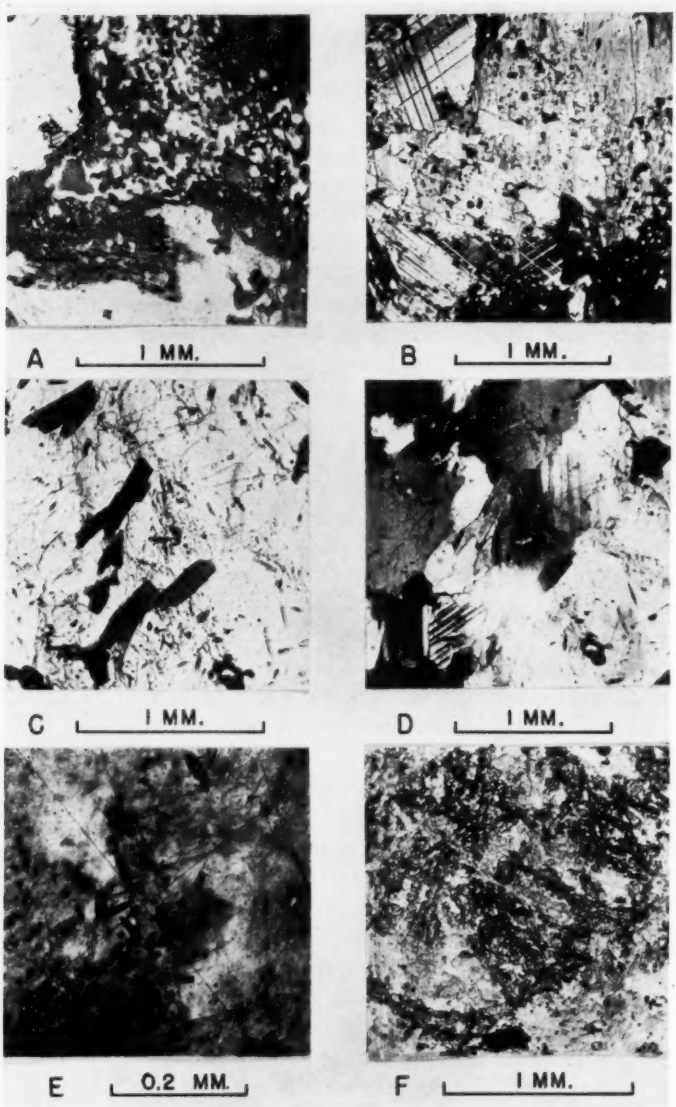

FIG. 7.—*A*, Ordinary light. Poikilitic hornblende with quartz chadacrysts. Gull Lake mass. *B*, Crossed nicols; same as *A*. *C*, Ordinary light. Hornblende grains and needles replacing andesine-labradorite (An_{50}). From south of Gull Lake. *D*, Nicols crossed; same field as *C*. *E*, Nicols crossed. Quartz grains and hornblende needles in andesine-labradorite. From south of Gull Lake. *F*, Ordinary light. Aggregates of epidote (dark gray), enclosing calcite, chlorite, quartz and acid plagioclase. The whole mass is interpreted as altered pyroxene crystal. From north contact.

habit in places.[17] Because of the lack of positive evidence, however, the quartz is regarded as older and as representing replacement blebs and grains in plagioclase that persisted when the plagioclase was replaced by hornblende.

An enormous number of quartz-diorite dikes cut the labradorite-hornblende rock, and near them and near the edge of the mass there is a marked development of epidote, clinozoisite, and zoisite and chlorite. The constituents are in ragged grains replacing both hornblende and labradorite, and there are also a few irregular veinlets cutting the rock.

These changes are complete in the eastern mass south of Gull Lake, but all gradations from this type to the Bell River type can be found in the area south of Olga Lake. These changes involved an addition of silica (the quartz and hornblende) and of water in the hornblende. The alteration occurred when the rock had solidified and does not at all resemble the deuteric changes (serpentinization) in the western part. The rock is also riddled by quartz-diorite dikes and apophyses as well as quartz veins.

The conclusion drawn, therefore, is that solutions from the quartz-diorite magma, while it was still far from the complex, flooded the rock and caused the pronounced development of quartz and hornblende. Later, cooler solutions from the same source caused the epidotization and chloritization observed near the dikes, and still later, cooler solutions were responsible for the numerous quartz veins.

SUMMARY ACCOUNT OF THE HISTORY OF THE COMPLEX

The complex, as stated above, is probably concordant and has been folded into a tight syncline with vertical dips on the limbs. A lopolith forms by upwelling of magma and settling of the floor so that the intrusion has a general saucer shape.[18] When the Bell River lopolith started to form, the volcanic complex and sediments were probably not much, if at all, deformed. Later movements have compressed both the country rocks and the complex into tight synclines. This folding was probably accompanied by the rise of the quartz-

[17] Freeman, *op. cit.*, p. 693.

[18] F. F. Grout, "The Lopolith: An Igneous Form Exemplified by the Duluth Gabbro," *Amer. Jour. Sci.*, Vol. XXXXVI (1918), p. 520.

diorite magma, the formation of migmatites, and the emplacement of the batholiths. This sequence of events is outlined in the article describing the dioritic batholiths of the area.[19] How much of this folding occurred before the final consolidation of the complex is not determined. The remarkable banding at the Mignon and Seven Channel rapids may have been due to differentiation helped by escape of mineralizers, which in turn may have been due to movements in the partly consolidated magma. These movements may have been caused by settling of the floor or may be due to active compression of the whole mass by orogenic forces. The remarkable amount of cataclastic texture in the upper part of the mass (the central part of the anorthosite area shown in Fig. 1) probably is due to much more vigorous movements in the crystal mush formed in the partly consolidated magma than mere basining of the mass through crustal settling. It seems more probable that active orogenic movements compressing the crystallizing magma into a syncline were necessary.

Also, at a relatively early stage deuteric, or hypohydrous, processes became active and caused the serpentinization of the pyroxene in parts of the lopolith. Since deuteric changes occur in a closed system, serpentinization is widespread, not related to zones of weakness or to intrusive rock.

The next event was the flooding of much of the rock of the complex with magmatic solutions from the rising quartz-diorite magma and the formation of the hornblende rock in the eastern part. These solutions not only caused a profound change there but also acted on much of the rest of the complex, causing less profound changes.

Finally, dikes and apophyses of quartz diorite were injected into the complex, and solutions accompanying and preceding them caused extensive chloritization and epidotization. The magmatic effects then ceased, and long-continued erosion stripped off the cover of overlying rocks to expose the present surface.

[19] Freeman, *op. cit.*

"LYELL" GULLY, A RECORD OF A CENTURY OF EROSION

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ABSTRACT

Sir Charles Lyell, when he visited the United States in 1846, published a description and a woodcut of a gully which he observed near Milledgeville, Georgia. Several photographs and statements give subsequent information on the gully since 1846. Similar gullies are widespread in the Piedmont area, many of which have been studied by the writer. An accurate survey and detailed study of the gully using physiographic evidence, the age of old pine trees in the gully, and the details of Lyell's description make it possible to reconstruct the conditions of 1846 and to interpret the changes since that date. As the gully was 20 years old when Lyell saw it, there is thus a documented history of over 110 years of erosion.

In 1846 Sir Charles Lyell, during his second visit to the United States, observed a large actively eroding gully near Milledgeville, Georgia. The dimensions of the gully as estimated by Lyell and others[2] are far from correct, and no description based on actual measurements has been found. However, since the history of the gully can be made out to a fair degree of accuracy from existing records, it has been considered worth while to make an accurate survey and detailed study of the area to provide a record for future use. Recent studies of the rate and history of gully erosion in the Piedmont have been based, in part, on information secured from the older residents in the vicinity, but such information rarely extends back more than 50 years. The documented history of "Lyell" gully, in contrast, covers a period of more than 110 years. Such a record is of especial value in the study of erosion because thousands of gullies of the same type, caused by accelerated erosion induced by man's occupancy, are found throughout the Piedmont. In order to study the rate and history of gully erosion, it is necessary to have a knowledge of former conditions to serve as a basis for comparison with the observable physiographic and other types of evidence.

[1] Associate Geologist, Section of Climatic and Physiographic Research, Soil Conservation Service. Published by permission of the Soil Conservation Service.

[2] Ulrich Bonnell Phillips, *Life and Labor in the Old South* (Boston: Little, Brown & Co., 1928), p. 11; F. Grave Morris, "Soil Erosion in the Southeastern United States," *Geog. Jour.*, Vol. XC (1937), p. 367.

Lyell's description is as follows:

The surprising depth of some of the modern ravines in the neighborhood of Milledgeville, suggests matters of curious speculation. At the distance of three and a half miles due west of the town on the direct road to Macon, on the farm of Pomona, is the ravine represented in the annexed woodcut. . . . . Twenty years ago it had no existence; but when the trees of the forest were cut down, cracks three feet deep were caused by the sun's heat in the clay; and, during the rains, a sudden rush of water through these cracks, caused them to deepen in their lower extremities, from whence the excavating power moved backward, till, in the course of twenty years, a chasm, measuring not less than 55 feet in depth, 300 yards in length, and varying in width from 20 to 180 feet, was the result. . . . . The high road has been several times turned to avoid this cavity, the enlargement of which is still proceeding, and the old line of the road may be seen to have held its course directly over what is now the widest part of the ravine. In the perpendicular walls of this great chasm appear beds of clay and sand, red, white, yellow, and green, produced by the decomposition in situ of hornblendic gneiss, with the layers and veins of quartz and feldspar, which remain entire to prove that the whole mass was once crystalline.

In another place I saw a bridge thrown over a recently formed gully, and here, as in Alabama, the new system of valleys and drainage, attendant on the clearing away of the woods, is a source of serious inconvenience and loss.

I infer from the rapidity of the denudation caused by running water after the removal or clearing of the wood, that this country has always been covered with a dense forest, from the remote time when it first emerged from the sea. However long may have been required to raise the marine tertiary strata to the height of more than 600 feet, we may conclude that the surface has been protected by more than a mere covering of herbage from the effects of the sudden flowing off of the rain water. . . . .[3]

The woodcut mentioned in this quotation is reproduced here as Figure 1. The illustration, combined with Lyell's description of the gully, forms an excellent basis for comparison with the present conditions.

In 1902 Thomas L. Watson,[4] of the Georgia Geological Survey visited the "Lyell" gully. Although a search has revealed no written description of the gully, two of his photographs are available and are

[3] Sir Charles Lyell, *A Second Visit to the United States of North America* (London: Murray; New York: Harper & Bros., 1849), Vol. II, pp. 23–25. The writer is indebted to Mr. Arthur Hall, of the Soil Conservation Service, for calling attention to this passage by Lyell.

[4] "Granites and Gneisses of Georgia," *Georgia Geol. Surv., Bull. 9-A* (1902), Pl. XXXI on p. 320.

Ravine on the farm of Pomona, near Milledgeville, Georgia, as it appeared January, 1846.

Excavated in twenty years, 55 feet deep, and 180 feet broad.

FIG. 1.—The woodcut of Sir Charles Lyell

reproduced as Figures 2 and 3. Several photographs were also taken in 1922 by S. W. McCallie, of the Georgia Geological Survey, but again no notes concerning them have been found. Only isolated portions of the gully were photographed; consequently, a complete comparison between present conditions and those that existed in 1922 is not possible. In 1937 F. Grave Morris,[5] a fellow-countryman of Lyell, published an article that included a splendid description of the gully and one photograph, but the dimensions given are estimates rather than actual measurements.

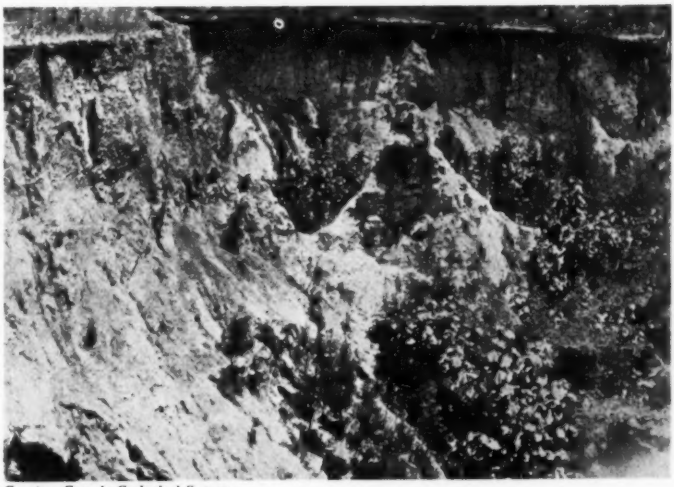
Today the gully is known locally as the Hall gully and takes its name from the present owner of the land, Dr. T. M. Hall, of Milledgeville, Georgia, who is widely known as an ornithologist and for the bird sanctuary surrounding his residence. Investigation shows that the gully was not located on the Pomona farm, as described by Lyell, but on the property of Captain Samuel Beckcom, a few hundred feet east of the Pomona boundary. There is no question of identity since there are no gullies today comparable in size to that described by Lyell on what was formerly the Pomona farm. Dr. Hall's grandfather, Iverson L. Harris, owned the Pomona farm at the time of Lyell's visit. W. H. Hall later purchased the adjoining Beckcom property and both tracts are now owned by his son, Dr. T. M. Hall. The grave of Captain Beckcom (1760–1825) is located about 60 feet from the present gully head and will eventually be undermined by the headward cutting of the gully. The grave is covered with a marble crypt and is known locally as the Revolutionary War Hero's grave.

The gully is located just as Lyell described it, about $3\frac{1}{2}$ miles west, but slightly southwest, of Milledgeville on the old Macon or Garrison road now known as the lower Macon road. A long narrow ridge 250 feet high extends to the eastward between the drainage basins of Fishing and Camp creeks, which drain into the Oconee River. The ridge is several miles long and from a few hundred feet to one-half mile in width. The area is representative of the topography at the contact of the crystalline rocks of the Piedmont and the sediments of the Coastal Plain. The gully has been cut into the shoulder of the steep slope on the south side of the ridge, a few hundred yards east of

[5] *Op. cit.*

Courtesy Georgia Geological Survey

FIG. 2.—View northeast across the heads of basins *B*, *C*, and *D* of the main gully in 1902. Photograph by Thomas L. Watson (cf. Fig. 4).



Courtesy Georgia Geological Survey

FIG. 3.—View northwest across basin *A* in 1902, approximately the same as Figure 6. Photograph by Thomas L. Watson (cf. Fig. 4).

a crossroad that is situated about 2 miles west of the Sanatorium. The area above the head of the gully has a slope of about 2 per cent. The main portion of the gullied area is on a 10 per cent slope, which increases to 40 per cent lower down, where the shoulder descends to the level of the major valley. At present the gully is 750 feet long, 500 feet wide, at the maximum, and 280 feet wide at the head, where vertical drops of 45 feet occur. The greatest depth of the gully is 60 feet near the central portion, and from the level of the rim at the head to the exit of drainage there is a fall of 111 feet. The grade of the gully floor is $2\frac{1}{2}$ per cent in the lower channel and averages 30 per cent in the upper channels, not including the vertical caveheads. In its lower reaches the gully is U-shaped with a flat floor aggraded by debris washed from the active heads above. The walls are abrupt, but are vertical only near the top because of the accumulation of talus debris at the base. The upper part of the gully system has vertical steep walls with slumped and caved material at the base, most of which is washed out during rains (see Fig. 6). The upper segment of the gully is compound, covering 2.12 acres and is shown on the map (Fig. 4) as drainage basins *B*, *C*, and *D*. The western tributary, basin *A*, covers 0.66 acres; the northeastern tributary, basin *E*, 0.53 acres; and the lower segment below the juncture of the tributaries, 0.55 acres. The total area occupied by the gully is 3.86 acres.

The gully has been eroded through the soil, subsoil, and deeply decayed, friable, crystalline rocks of the Piedmont Plateau. The crystalline rock on the upper slope is overlain by a thin layer of quartz conglomerate tenaciously cemented by iron oxide. The conglomerate represents the basal deposit of the Tuscaloosa formation of Cretaceous age. The soil is Cecil clay loam, derived from a fine-textured gneiss. The subsoil is several feet thick, very clayey and tightly bonded, red in color, and underlain by the decayed parent material or saprolite more than 60 feet in depth. The fine-grained and clayey character of some portions of the parent material is responsible for many of the slumped areas along the gully rim (see letter *S* on Fig. 4). The largest block, of about 1,800 cubic yards, collapsed in 1930 and disturbed the drainage to a considerable extent. Conditions at the base of the slump indicate that the mass still

moves, occasionally. When large volumes of the material are removed from the front of the mass by run-off the block moves forward and will continue until equilibrium is established.

Because of the character of the soil profile in the gullied area, the process of head-cutting is significant. When run-off concentrated in a channel falls over the rim of the gully, a plunge pool develops at the bottom. Some of the water at the lip is diverted as a thin film down along the face of the exposed resistant material of the subsoil by the action of surface tension. The water, upon reaching the underlying friable parent material, is absorbed, and, as saturation occurs, the material falls to the floor of the gully as mud. Spray from the waterfall or splashing from the plunge pool may contribute additional moisture to the walls. By these processes the parent material recedes beneath the overlying resistant layer, and a cave develops. Eventually, the cave becomes so recessed that the overhanging subsoil collapses. Thus the gully head advances upslope as the process of undermining is repeated. The height of the cave-head is limited only by the depth of the decayed parent material or by the grade level of the gully.

The development of the cave and its collapse may require several months, or it may occur two or three times during one rainstorm. Gully-cutting is more rapid during rains of high intensity. The run-off sweeps out the fallen material from the gully floor, thereby assisting in the renewal of conditions favorable to cave-head enlargement. A deep plunge pool is not a clearly observed feature in gully erosion in the Piedmont, since it is rapidly filled with debris during a rain and remains filled by the debris washed into it as the run-off decreases. The action of surface tension continues as long as there is any water seeping over the rim of the gully, even after the free flow from run-off has ceased. Consequently, erosion by means of the caving process continues long after the rain has stopped, but the accumulated caved material is not likely to be removed until the following rain.

Long low-intensity rains saturate the rim and walls of the gully causing them to cave, but the run-off is seldom sufficient to carry away the fallen material. A continued lack of removal or a reduction in the amount of run-off caused by a decrease in size of the drainage

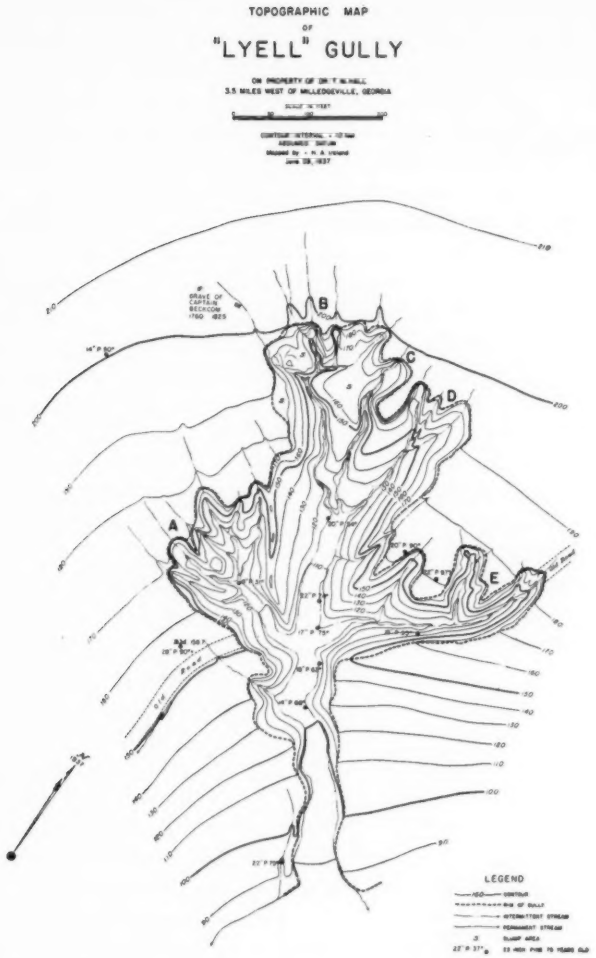


FIG. 4.—Topographic map of "Lyell" gully on the property of Dr. T. M. Hall, Milledgeville, Georgia, June, 1937.

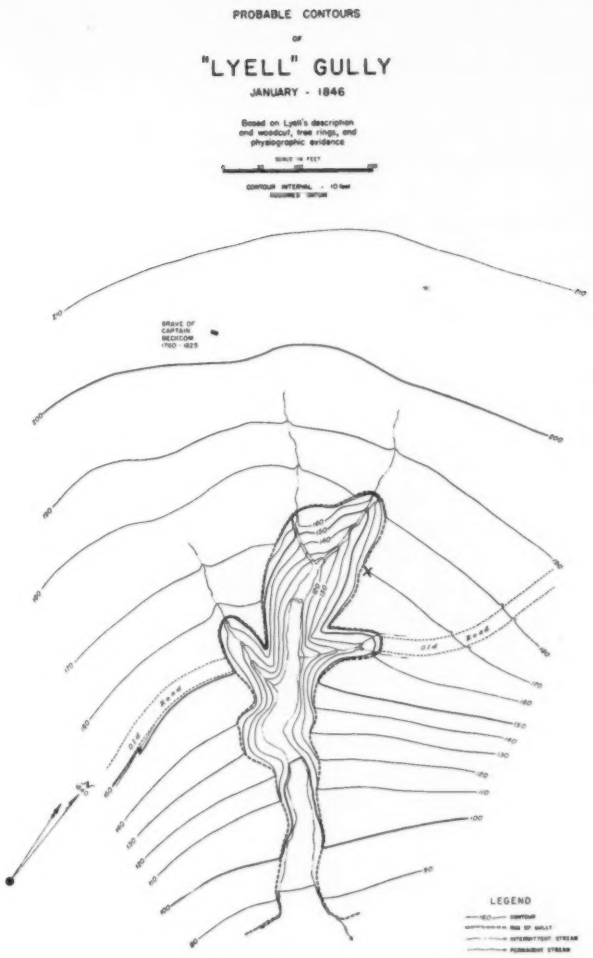


FIG. 5.—Probable contours of "Lyell" gully, January, 1846, based on Lyell's description and woodcut, annular growth of tree rings, and physiographic evidence.

basin will retard the formation of cave-heads. If the collapsed debris accumulates above the top of the friable saprolite, it may prevent further head recession. Eventually, the material will form a slope, and natural stabilization will occur as vegetation has a chance to grow without being washed away or covered. Where these conditions exist there may be a large number of actively eroding heads in a single gully system. Associated with them may be several retarded or stabilized heads. In the "Lyell" gully system there are fifteen active cave-heads, having vertical drops of 15–45 feet, and four that have been stabilized.

Lyell states that the gully was nonexistent 20 years before he visited it in 1846. The gully, accordingly, has been enlarged to its present size in less than 112 years. Field evidence indicates that there was no large well-established drainage channel on the slope to localize the gullying processes. There are no steeply sloping shoulders of valley sides, which should survive if a former drainage channel of large size existed. The accordant levels and abrupt rims on opposite sides of the gully, especially in the lower segment, show that the hillside was unbroken by any pre-existing channel. The head of the gully lies in a broad dalelike curve on the side of the slope (see Fig. 4), where the run-off was dispersed by the forest litter instead of following definite channels. Later, the water was concentrated into channels and the gully initiated. The start of the gully is attributed to a drainage channel leading from the ditch of the old road. This was enlarged by increased run-off from the slopes above, after they were cleared and put into cultivation. The gully at present is enlarging at a less rapid rate on the steeper portions of the hillside than it was in its early stages, because the drainage area tributary to the gully head has been reduced in size as the gully advanced headward into the upland. In addition, the present upland drainage area is partly covered with a second-growth wood that tends to reduce the run-off. The gully head has now reached the more gently sloping land near the top of the ridge.

The accompanying map (Fig. 5) shows the probable outline and character of the gully in 1846 and is based on Lyell's description and woodcut, on the age of the trees in and around the gully, and on physiographic evidence. The head of the gully in 1846 was probably

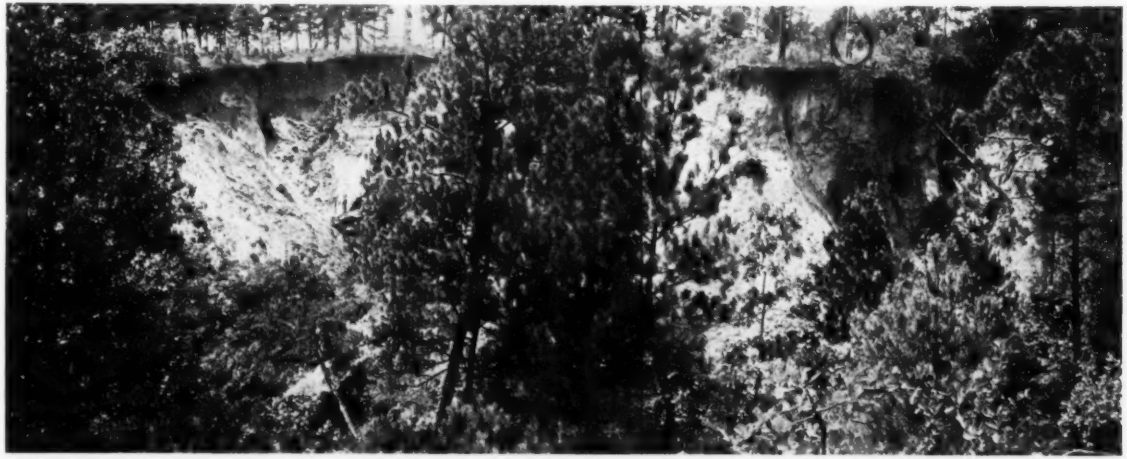


FIG. 6.—The actively eroding heads of basin *B* are at the left and of basin *D* at the right. Captain Beckcom's grave is just above basin *B*. Photo taken in the opposite direction from that shown in Figure 7.

located where the upper part of the present main channel narrows and receives the drainage of basins *B*, *C*, and *D* (cf. Figs. 4 and 5).

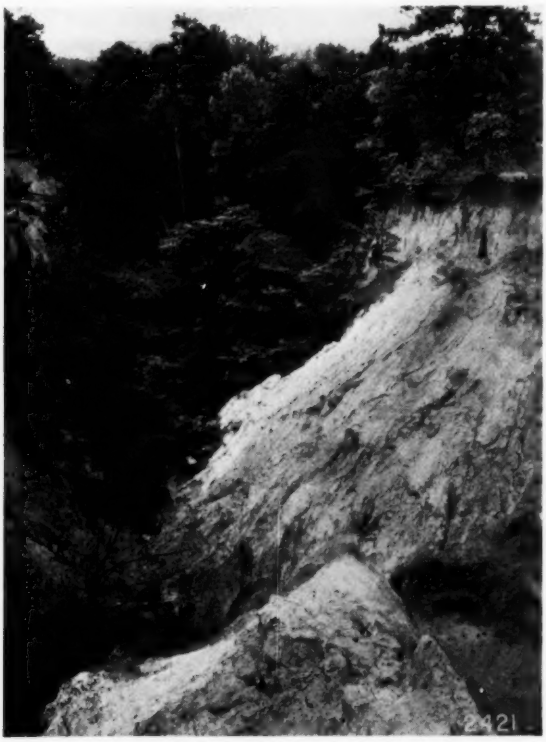


FIG. 7.—Looking southeastward down the main channel of the gully. Note the slumped block at the right and the headward margin of vegetation near the center of the view. The gully is deepest below the bare walls shown in the left background. The ridge in the far distance is the same as that in Lyell's woodcut.

Observations on several dozen active gullies in other parts of the Piedmont indicate that the boundary or margin between advancing vegetation and the exposed nonvegetated upper part of an actively eroding gully generally lies 100–200 feet downstream from the cave-heads (Fig. 7). The growth of permanent vegetation is one of the

processes of natural stabilization, but it will not occur until the activity of erosion is reduced, or until the point of activity has moved far enough upslope to allow seed to germinate and the young plants to grow without being covered or washed away.

The increment boring of a pine tree in the bottom of the gully, just above the position of the old road crossing, shows that the tree is 75 years old, growth having started in 1862. Ten years is allowed to give time for the development of conditions favorable for the young pine to start its growth. The allowance is based on the rate of cave-head recession as 10–20 feet per year. This would place that portion of the gully as being the graded and less disturbed bottom in 1852, and the probable margin of vegetation at that time. The active head of the gully would, therefore, have been about 100–200 feet upslope and slightly farther up than the position of the rim as viewed by Lyell. In this way the position of the gully head for 1846 as shown in Figure 5 was determined. The probable location of the point where the two figures shown in the woodcut are standing is marked by a cross on the map.

As the actively eroding areas migrated upslope, the vegetation followed and reached the juncture of basins *B*, *C*, and *D* about 1883, according to the increment boring of a 54-year-old pine tree located at that point. The trunks of the bored trees were straight and uniform, and the increment core showed the heart of each tree essentially at the center. Using an allowance of 10 years, 1873 would be the date at which that portion of the gully became the graded bottom. Since then vegetation has advanced slowly into basin *B* (Fig. 8), being retarded by the repeated slumping of the clayey decayed gneiss and the continued action of the run-off from the fields above.

The huge block of 1,800 cubic yards which slumped into basin *B* in 1930 carried with it several pine trees over 40 years old. The upper portions of these trees are now curved to a vertical position from the former inclined attitude. This block formed a part of the rim between basins *B* and *D*, and if restored would show the rim as it was in 1930. The indentation left in the rim now has a cave-head 25 feet high and the drainage from it has cut basin *C* down through the slumped material. Formerly all of basin *C* drained through basin *B*. The most eastern branch of the upper channel, basin *D*, shows a

normal advance of vegetation. The advance has been more rapid than in the others, as only one of the three heads is active. Pine trees 10–30 years old are found in the upper and central part, both in the channel and along the sides, but especially on the east side. In some places the vegetation in the gully is continuous with that on the rim and is so dense that it is not possible to get to the rim. The vegeta-

FIG. 8.—The cave-heads are typical of those developed in the Piedmont of southeastern United States. View west showing the active multiple heads in basin *A*.

tion throughout the gully is second-growth pine with a thin to dense undergrowth of scrub oak, plum, sweet gum, shrubs, and briars.

The maximum depth of the entire gully, taken from the northeastern rim of the main channel near the position of the old road, to the bottom of the gully, is 60 feet. Lyell's observation of a depth of 55 feet was probably made at the point of present maximum depth. Lyell estimated the length of the ravine as 900 feet but the present measured length is only 750 feet. The length as reconstructed for 1846 is 600 feet. Time did not permit Lyell to make detailed measurement of everything that he observed. An estimate of the length was in all probability hindered by a wood which obscured the

mouth of the gully. That there was such a wood is indicated by the age of the present trees and by the trees shown in his illustration. Although not correct, the estimated dimensions serve admirably to illustrate the magnitude of erosion as judged by a very observant geologist.

Observations have shown that roads and road ditches are among the chief causes of gully-cutting because of the concentration of run-off and the lack of vegetation in them. Throughout the Piedmont roads have had to be abandoned or relocated because of erosion. The position of the road described by the famous geologist is shown by an embankment about a foot high, located in the area marked "old road" south of the rim of basin *A*, and by slight depressions of the old ruts east of basin *E*. Large pine trees, some over 90 years old, indicate that the area has not been cultivated since 1847 at the latest. Lyell states very definitely that the old road had been washed away by the gully erosion and that the widest part of the ravine was where the old road originally crossed. The embankment is aligned across the widest part of the present gully and is opposite the mouth of the northeastern tributary, designated as basin *E*. This relation is evidence that the basin was caused by the old road (see Fig. 5). The trees in the right background of Lyell's sketch are interpreted as being along the old roadside.

The evidence discussed leads to the conclusion that the main head of the gully in 1846 was at the position indicated on the map of reconstructed conditions for that date. The head of the gully was above the old road and was less than 750 feet from the mouth. The gully was deepest at a point which has the maximum depth today. The characteristics of the vegetation also support the conclusion. The position of the head could not have been farther downslope since the depth becomes rapidly less, much below 55 feet, and the length of the gully would be less than 600 feet and too small to fit the length of 900 feet as given by Lyell. Location of the head farther upslope was not likely as more than 150 feet of erosion to the present head would probably have occurred in the 90 or more years since 1846, especially since about 600 feet of gullying occurred in the 20 years before Lyell saw the area.

The retreat of the gully heads is slower now than formerly (Fig. 9), because the heads are within 200–300 feet of the drainage divide and receive much less water than they did previously. The area north of the gully is still cultivated and planted to cotton. The area to the west was abandoned about 1880 and now is covered with a second-growth stand of pine over 50 years old. Since 1880 the vegetation has retarded the run-off, and consequently the gullying has become much less rapid, although it is still fairly active (Fig. 8). The cultivation of the area west of the gully was responsible for the development of basin *A* and part of basin *B*. Small pine and other pioneer plants are now growing in the channel of basin *B* to within

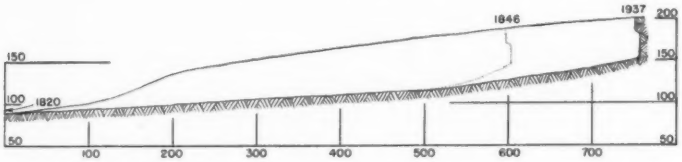


FIG. 9.—Profile of "Lyell" gully showing the stages of cutting

50 feet of the heads. The road, as previously stated, concentrated the water which cut basin *E*. A pine tree on the north side of this basin was 6 years old at the time of Lyell's visit and shows that the area has not been in cultivation since about 1840. Cultivation of the slopes above the tree was responsible for the development of the upslope branches of basin *E*. The age of other trees indicate that that area has not been plowed for over 40 years. The heads, though active, have been much retarded by the growth of a dense undergrowth, which occupies the upslope area at present. The caved material that has accumulated in the gully heads is not being removed, and natural stabilization will probably be established within the next few years. Vegetation in the bottom of the tributary is now fairly well advanced.

Phillips writes concerning the "Lyell" gully:

> When I last visited the chasm, fourscore years after Lyell, its length and breadth had perhaps doubled. But great pines had fallen with the caving banks, taken root afresh with trunks aslant at strange angles, and were holding the floor as a fairly permanent slope. The gully's head is now so near the hill

crest that its advance is slackened; and its threat of engulfing the lonely grave of a revolutionary veteran and the slab placed thereon by the state may not be accomplished for several score years to come.[6]

Even though it is slower than formerly, the rate of gully erosion is such that the "engulfment" of Captain Beckcom's grave is more likely to occur in the next two or three decades than in several score years. An old negro living near the gully who has been watching its retreat toward the grave asked one group of observers if they had come to remove the grave to another location.

Gullying is directly traceable to the utilization of the land by man. The land when improperly used may be eroded disastrously as the thousands of gullies in the Piedmont show. "Lyell" gully is one of the oldest of these and has the most complete recorded history. Even where the land is well used, if the drainage is not controlled properly, accelerated erosion by concentrated run-off presents a hazard. Lyell, nearly a century ago, recognized some of the effects of man's ultilization of the land when he wrote:

> Here [Columbus, Georgia], as at Milledgeville, the clearing away of the woods where these Creek Indians once pursued their game, has caused the soil, previously level and unbroken, to be cut by torrents, so that deep gullies may everywhere be seen; and I am assured that a large proportion of the fish, formerly so abundant in the Chattahoochie, have been stifled in the mud.[7]

Today, the demands on the soil by a population many times that of 1846 makes the hazard of soil erosion a problem which must be kept in mind.

ACKNOWLEDGMENT.—The writer is indebted to Mr. O. D. Price for assistance in the field work.

[6] *Op. cit.*, p. 11.

[7] *Op. cit.*, p. 33.

SOME TYPES OF BEDDING IN THE COLORADO RIVER DELTA[1]

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ABSTRACT

The subaerial portion of the Colorado River Delta may be divided on the basis of texture and structure into three principal areas. The upstream area is one of deposition during periods of low-river stage only, and its deposits are almost exclusively fine sand, showing in vertical section intricate patterns of small-scale cross lamination formed by current ripples. The middle area is the principal site of deposition today and is characterized by many local deltas built up on the general plain. During low-river stages mud alone is deposited here, but with higher waters the mud is carried beyond and fine sand is deposited. The mud is structureless or flat bedded; the sand shows cross lamination either of the ripple variety or of a type comparable to the foreset beds of a simple experimental delta. The lower part of the delta, including the estuary, is entirely of mud today due to recent severance of direct connection between the river and the sea. The deposits are uniformly flat bedded except where erosion channels have been scoured out by retreating tide waters and later filled with layers conforming to the contours.

The principal types of lamination on the delta are described and classified and two varieties of pseudobedding are discussed in this paper.

INTRODUCTION

Many papers have been written on the development of deltas, especially from the standpoint of geographical features and from that of composition and growth. The studies of Gilbert[2] and Barrell[3] on delta development are classic, and the recent work of Trowbridge[4] on the Mississippi Delta is outstanding. Interesting experimental studies in this field have been conducted by Smith[5] and by

[1] Published with permission of the Carnegie Institution.

[2] G. K. Gilbert, "Lake Bonneville," *U.S. Geol. Surv. Mon. 1* (1890), pp. 153–67; "Topographic Features of Lake Shores," *5th Ann. Rept. U.S. Geol. Surv.* (1885), pp. 104–8.

[3] Joseph Barrell, "Criteria for the Recognition of Delta Deposits," *Bull. Geol. Soc. Amer.*, Vol. XXIII (1912), pp. 377–446; "The Upper Devonian Delta of the Appalachian Geosyncline," *Amer. Jour. Sci.*, Vol. XXXVI (1913), pp. 429–72; *ibid.*, Vol. XXXVII (1914), pp. 87–109, 225–53.

[4] A. C. Trowbridge, "Building of the Mississippi Delta," *Amer. Assoc. Petrol. Geol. Bull.*, Vol. XIV (1930), pp. 867–901.

[5] A. L. Smith, "Delta Experiments," *Amer. Geog. Soc. Bull.*, Vol. XLI (1909), pp. 729–42.

Nevin.[6] Despite the wealth of material on the general subject, however, the writer has been unable to find in the literature data affording criteria for the identification of delta structure in geologic formations.

During sedimentary studies of Permian formations in the Grand Canyon region, made by the writer in 1934, 1935, and 1936, certain types of cross lamination were found very puzzling. Explanation was sought in areas of present deposition where conditions comparable to those of the Permian are thought to exist. The delta of the Colorado River, in Mexico, where fluviatile, aeolian, and marine sediments merge and where an arid climate prevails, seemed to the writer to promise an answer to some of the problems presented by the Permian sediments.

CHARACTER OF THE DELTA

The general features and character of the Colorado River Delta are well known today through the recent publications of Sykes[7] which summarize the results of forty years of observation and study in this region. The subaerial portion of the delta covers an area of approximately 3,325 square miles, most of which is in Mexico with the exception of one large northwestward-extending arm which fills the Imperial Valley of California. Sykes compares the shape with that of a letter T having a broad short stem formed by the portion southwest of Yuma, Arizona, where the river begins its deposition. The full spread of the two arms in this figure is nearly 200 miles, and the major dimension perpendicular to this is approximately 70 miles.

The nature of the subaqueous portion of the Colorado Delta is known only in a general way. For the first 200 miles the deposits slope seaward under the waters of the Gulf of California about 15 feet to the mile, then comes a marked increase in grade, next a relatively flat surface, and finally another steep slope reaching a depth of about 9,000 feet at the mouth of the gulf. Further discus-

[6] C. M. Nevin and D. W. Trainer, Jr., "Laboratory Study in Delta-building," *Bull. Geol. Soc. Amer.*, Vol. XXXVIII (1927), pp. 451–58.

[7] Godfrey Sykes, "The Colorado Delta," *Carnegie Inst. Washington Pub. 460* (1937); "Delta, Estuary, and Lower Portion of the Channel of the Colorado River, 1933 to 1935," *Carnegie Inst. Washington Pub. 480* (1937).

sion of this portion of the delta may be omitted here, however, for it is beyond the scope of the present study.

In the examination and classification of original structures on the subaerial portion of the Colorado Delta, it is important to keep in mind three fundamental factors which, as shown by Sykes,[8] have contributed materially to the upbuilding and shaping of the delta. These are (1) the comparatively narrow space available for deposition, (2) the form of the pre-existing marine basin toward and into which the river flow has been directed, and (3) the tremendous tides, up to 30 feet in height, at the river mouth, acting upon the seaward parts. Another element having a profound influence on the distribution of various types of bedding over the delta surface is the fact that there has been a practical severance of direct communication between the river and the sea for the last thirty years so that during this period essentially "no detrital matter of the heavier, coarser, bed-silt type, such as is incapable of being maintained in suspension more than momentarily by the current, has entered the Gulf from the river."[9]

CHARACTER AND TYPE OF DEPOSITION

Because the Colorado River has virtually cut off its own free communication with the sea during recent years, deposition of all but the finest material is now necessarily taking place in or above the central part of the subaerial portion of the delta. As a result, the areas of recent deposition between Yuma, Arizona, and Montague Island in the gulf, at the other extreme, may be roughly classified into three parts. These consist of a middle portion where most of the present deposition takes place and where local deltas are continually being superimposed on the main delta surface; of a lower portion where only sheet floods, subsurface waters, and tide waters are affecting the deposits; and, lastly, of an upper portion where deposition consists chiefly of the heavier, coarser types of deposits laid down during low stages of the river when there is not sufficient power to carry this sediment to the local deltas of the middle portion.

The three portions of the subaerial delta as defined in this paper

[8] *Pub. 460*, p. 155.

[9] *Ibid.*, p. 156.

are distinctive both in the manner in which each has been built up and in the texture of the sediments involved. Since the differences have been brought about, for the most part, by varying degrees of the same or related forces, the change from one type to another is necessarily transitional. The zone of gradation between the middle and lower portions of the delta is relatively narrow, for it represents the lower margin of modern deltaic fans which are developing on the subaerial surface of the main delta. In contrast, the gradation from upper to middle portion is so gradual as to be nearly imperceptible.

The middle portion of the region of deposition, as defined in this paper, includes a band roughly 15 miles wide extending across the delta from northwest to southeast surrounding and paralleling the Mexicali and Gulf Railway embankment. Information obtained from Sykes indicates that within this band the area of extensive deposition and local delta-building has in recent years migrated from northwest to southeast (Fig. 1). Periods of upbuilding in each of three major areas within this belt were 1909–23, 1923–29, and 1929–34.

The character of deposition in the middle portion of the subaerial delta is illustrated in the descriptions by Sykes of certain floods. Referring to the one of February, 1932, he states: "Vegetation was generally overwhelmed, and the invasion took the form of a sheet flooding over the whole surface of the cone, rather than the more common building up in detail along dispersal channels."[10] The latter type, he states, occurred in the flood of August, 1933, when "actual deposition of the tractional load of the current [was] practically confined . . . . to the distributaries and the innumerable small outlets which fringe the banks of the partially ponded main channel." A third type of deposition is that due to wastage of flood waters in paludal areas through seepage, evaporation, and absorption by vegetation. This takes place primarily during periods of moderate or minimum discharge, because in flood periods the fine suspended load is carried on to the lower portion of the subaerial delta and to the sea beyond.

From a textural point of view sediments in the middle portion of

[10] *Ibid.*, p. 159.

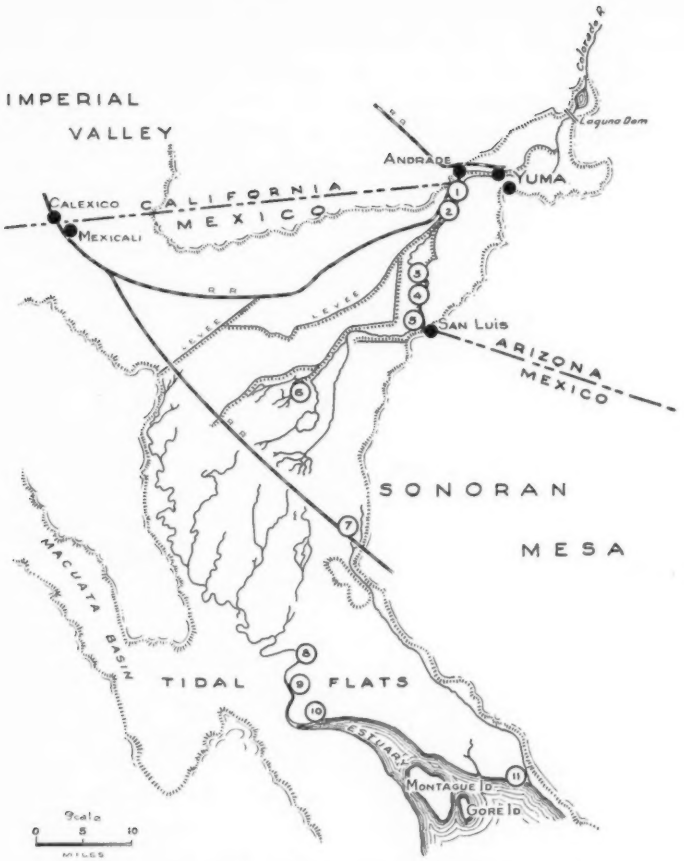


FIG. 1.—Map of Colorado River Delta (after Sykes). Principal localities examined indicated by numbers: *1–5*, Colorado River, Yuma to San Luis; *6*, Pescadero; *7*, one mile from east end of railroad; *8–10*, lower river and estuary; *11*, tidal flat near Sonoran Mesa.

the delta are of two types. One of these is mud[11] resulting from deposition of the suspended load in this area during low and moderate stages of the river; the other is fine sand which is introduced as tractional load during the high and moderate stages. The mud is uniformly fine and does not contain particles large enough to be retained in a 230-mesh screen. Typical samples of fine sand from three localities in this part of the delta, however, show a concentration in grain size around 0.2 mm. (retained by 120-mesh screen)

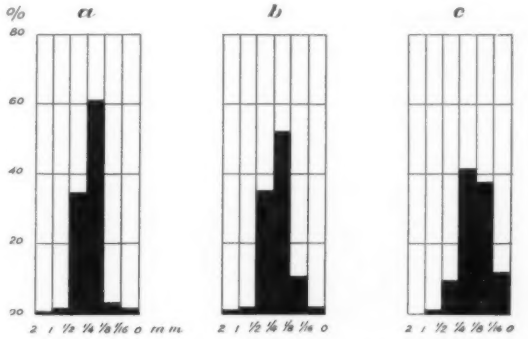


FIG. 2.—Histograms illustrating texture of typical sand deposits in middle portion of delta.

(see Fig. 2, *a*, *b*, *c*). These measurements agree closely with results of many analyses by Sykes of the tractional load of the river south of Yuma. He states that "about 85 per cent of the material represented . . . . existed in size between 60 and 300 mesh and the maximum percentage was slightly above the 120 mesh."[12]

In the upper portion of the delta from the vicinity of Yuma south at least as far as San Luis, a distance of 20 miles, recent deposition has been largely confined to a type which involves accumulation of the river's tractional load in distributaries and bordering outlets during low stages. Essentially no mud (except in canals) is deposited

[11] The term "mud" as used in this paper refers to detrital sediment in which the grains are less than $\frac{1}{16}$ mm. in diameter, so includes both silt and clay of Wentworth's classification.

[12] *Pub. 460*, p. 131.

here, and the fine sand, of homogeneous texture, resembles that deposited in great quantities during higher stages on the secondary deltaic fans of the middle portion of the delta.

Present deposition in the lower portion of the subaerial delta is in marked contrast with that in the upper. Since the practical severance of direct communication between the river and the sea confines the tractional load moved by river currents to the middle and upper portions, deposits in and around the estuary today are restricted to the fine material of a suspended load. This reaches it in times of flood and is arrested and precipitated through contact with saline waters or is gradually dropped on the surface because of seepage, evaporation, or other agencies. This portion of the delta, unlike those above, is largely devoid of plant life, no doubt because of high salinity resulting from tidal invasions and subsequent evaporation of sea water.

Among the extremely fine mud deposits of the estuary and bordering parts of the delta, the shells of mollusks are locally not uncommon. Gastropods which were collected at a locality about 17 miles north of Montague Island were identified by Mr. Horace G. Richards as *Helisoma trivolvis* (Say), *Marinula rhoadsi* Pilsbry, and *Cerithidea californica* Haldeman. The first two of these are said to be freshwater species and the last a brackish-water type. *Helisoma trivolvis* also was found in fine-sand deposits of the middle portion of the delta near the eastern end of the railroad embankment.

METHODS OF STUDY

Although the subaerial portion of the Colorado River Delta represents an area of deposition greater than 3,000 square miles, only a comparatively small part readily lends itself to the type of study proposed. For the purpose of recording the actual character of the beds of sand and mud involved, it was necessary to find steep, freshly cut banks on whose faces the individual laminae were visible either because of differential weathering or because of color or textural variation. Fortunately, such exposures are best developed along the present river channel which, therefore, makes available representative samples of bedding from one end to the other of the subaerial portion of the delta, a distance of about 80 miles.

In order that the true character of laminations be completely recorded, photographs, sketches, or both were made, showing three dimensions wherever possible. Only rarely were natural conditions favoring such complete observations to be found, but by the use of a very long knife it was discovered that vertical or horizontal faces could easily be developed, and by the careful use of a blower the laminae could be readily etched out in relief.

BEDDING IN UPPER PORTION OF SUBAERIAL DELTA

Those sediments which have been deposited during recent years in the upper portion of the Colorado Delta, from Yuma southward about 20 miles, consist nearly entirely of bedload or tractional material which river currents, in times of low stage, have been unable to transport farther toward the area of principal deposition in which deltaic fans are now forming. The sediment is almost exclusively a fine sand with uniform grain size averaging around 0.2 mm. In all sections examined along the river between Yuma and San Luis, it shows cross lamination which is characterized by intricacy of pattern and by fineness of detail. Individual laminations range between $\frac{1}{32}$ and $\frac{1}{8}$ inch in thickness and normally weather out in relief in dry sand because of slight textural variations throughout the deposits.

Cross lamination in deposits of the upper portion of the Colorado Delta is almost entirely the result of ripple development during deposition. The natural conditions under which various sets of ripple laminae form are many; therefore, resulting cross-lamination patterns in vertical sections are diverse. The character of these patterns depends primarily on whether or not successive series of rippled deposits are superimposed with the same rhythm and in the same direction. Furthermore, among the structures developed with uniform rippling, two types may be recognized—those in which sediment is uniformly distributed over the rippled surface and those in which there is concentration on the lee slopes. Finally, a factor which affects the appearance of all types of ripple laminations and completely controls that of ripple structures developed with lee-side concentration is variation due to deposition upslope, on a level, or downslope.

The following outline serves to classify the principal types of ripple cross lamination which are recognized on the Colorado River Delta:

1. Laminae superimposed in rhythm
   - A. Uniform deposition
   - B. Lee-side concentration
     - *a*) Upslope ripple
     - *b*) Downslope ripple
     - *c*) Level-surface ripple
2. Laminae superimposed out of rhythm

The simplest form of ripple cross lamination, and the only type in which the mode of origin is readily apparent, is that which in vertical section shows a series of uniformly superimposed asymmetrical ripple profiles (Fig. 3, *a*). These ripple outlines may be sloping up, or down, or may be essentially horizontal, depending on the original surface of deposition; however, in every case the form is easily recognizable. This type of ripple cross lamination is relatively scarce on the Colorado River Delta, for (1) most of the ripples are not of types which tend to be uniform in character (Pl. I, *B*), (2) vertical sections are only infrequently cut exactly normal to trough and crest, and (3) one set of ripples usually is not superimposed on another with identical direction and rhythm.

A type of ripple cross lamination which is consistent in form, but whose mode of origin is not usually obvious, is developed from successive rippled laminae being superimposed on each other with similar rhythm and direction but with deposition concentrated on the lee side of each ripple. Usually there is little or no deposition on the approach side. The result of this nonuniformity in distribution of sand is that a type of pseudobedding is developed by concentration or combining of laminae representing the approach slopes of the ripple deposits.[13] Each concentration develops into one of a series

[13] This pseudobedding resembles that figured and discussed by Gilbert ("Ripple-Marks and Cross-bedding," *Bull. Geol. Soc. Amer.*, Vol. X [1899], pp. 135–40) but differs from it in that the ripples involved are of asymmetrical, not of wave or symmetrical type. Pseudobedding described by Spurr (J. E. Spurr, "False Bedding in Stratified Drift Deposits," *Amer. Geol.*, Vol. XIII [1894], pp. 43–47), although also developed when the corresponding points in successive asymmetric ripple marks moved upward and onward with deposition, is entirely due to textural variation within ripple structures; therefore, it is fundamentally different from that described from the Colorado Delta.

of parallel lines, which dip in a direction opposite to that of actual ripple movement (Pl. I, *D*). Where the ripples ascend a sloping sur-

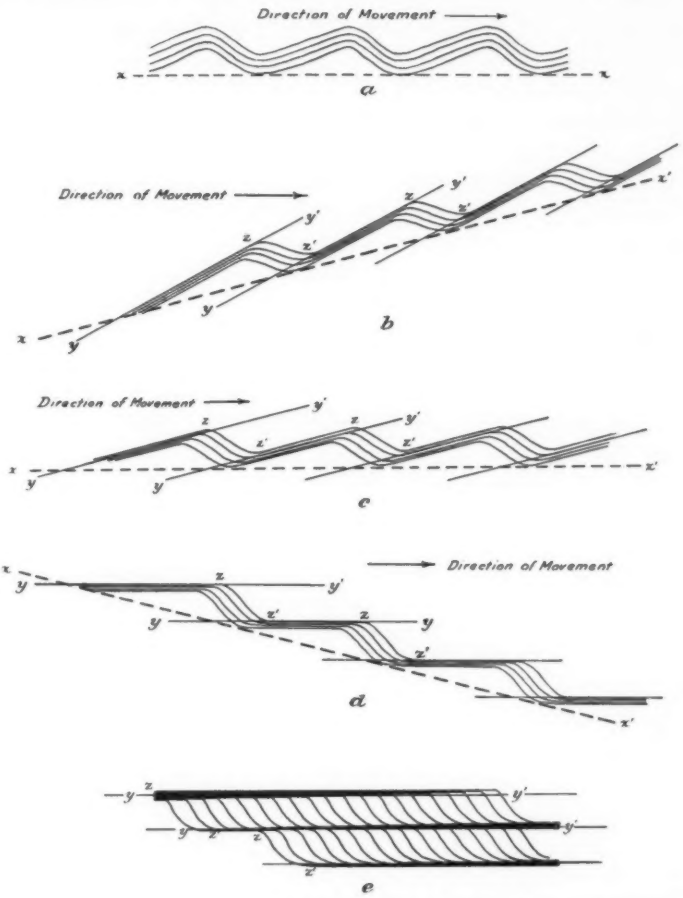


FIG. 3.—Cross lamination developed from ripples in rhythm: *a*, uniform deposition; *b*, *c*, *d*, lee-side concentration with deposition upslope, on level, and downslope; *e*, general effect of downslope ripple structure (*d*, when much repeated without change).

face the lines of pseudobedding dip steeply away, and the laminations formed from lee-side deposition approach a horizontal attitude. On

the other hand, where deposition is downslope, the pseudobedding lines approach horizontality and each of the lee-side laminae dips steeply. An intermediate situation, where ripples are deposited on a level surface, is characterized by gentle dips, but in opposite directions, of both pseudobedding lines and lee-side laminae (see Fig. 3, *b*, *c*, and *d*).

The extent to which a lee-side ripple deposit of any one type continues to be superimposed on another of the same type affects to a marked extent the appearance of the cross lamination as seen in a vertical section. If a small group of only six to twelve similar laminae is deposited, the general appearance in cross section is like a series of erosion pockets filled with laminae, and in many places it superficially resembles compound foreset bedding, as described by Lahee,[14] or even the festoon cross lamination of Knight,[15] as seen in a vertical plane normal to the troughs. On the other hand, where lee-side ripple cross laminae are formed continuously and without change of form for a considerable time, the resulting pattern in cross section will consist of apparent layers, and within each of these many nearly parallel laminae dip, forming tangents with the base. In the case of downslope ripples, the apparent layers are nearly horizontal so the dipping laminae within suggest torrential cross bedding, although on a very small scale (Fig. 3, *e*). In upslope ripples the laminae give the deceiving appearance of horizontal beds which have been eroded asymmetrically.

Despite the apparent similarity when seen in vertical section between lee-side ripple cross laminae and certain well-known types, such as compound cross laminae, the actual differences are great. The ripple type is developed through deposition alone; the other

[14] F. H. Lahee, *Field Geology* (New York: McGraw-Hill Book Co., 1923), p. 80. This type of cross-bedding in which each set of laminae is truncated by the next overlying set, is due to interference among adjacent lobes on a delta front. It should not be confused with "compound cross-bedding" which is defined by Gilbert ("Ripple-Marks and Cross-bedding," *op. cit.*, p. 139) as lamination structures "produced by deposition and partial erosion associated with shifting sand ripples" and which apparently resembles, except for planes of partial erosion, the type of ripple lamination pattern found on the Colorado Delta.

[15] S. H. Knight, "The Fountain and the Casper Formations of the Laramie Basin," *Univ. Wyo. Pub. in Sci.: Geology*, Vol. I (1929), pp. 56–74.

varieties involve both deposition and nearly contemporaneous partial erosion. Analyses in three dimensions readily show the true character of the ripple laminae and consequently, where practical, offer an excellent method for recognition. A section on the horizontal plane invariably illustrates that no erosion and only deposition is involved in the development of this structure, for laminations as seen on this plane are comparable to the contours on a topographic map—they may converge until they meet but will never cut across one another. This is not the case among those in scour-and-fill deposits. Typical samples, illustrated in three dimensions, are shown in Figure 4, *a*, *b*, and *c*.

Although the various definite types of ripple cross lamination described in the preceding paragraphs are recognized locally throughout the upper portion of the Colorado River Delta, most of the cross lamination here represents combinations of various types and sizes of ripples, and therefore its structure is extremely variable. Examination of rippled surfaces along the river today (Pl. I, *A*, *B*) shows that, for the most part, the troughs and crests are very irregular in form and have no constant direction of strike. There is little probability, therefore, that a superimposed ripple deposit will exactly conform to the underlying ripple form. Furthermore, irregular shape in most ripples of this type causes all semblance of uniform, rhythmic character to be lost in the laminae of vertical sections which are cut in directions other than that of movement. This is illustrated by a series of profiles of recent ripple surfaces made in various directions (Fig. 5). Thus, recognition of the mode of origin of much ripple cross lamination is not readily apparent; however, three-dimension analysis as previously indicated will invariably serve to give its identity.

BEDDING IN THE MIDDLE PORTION OF SUBAERIAL DELTA

In the middle portion of the subaerial part of the Colorado Delta, where most of the active deposition has been going on within recent years, small deltaic cones are developed with great rapidity. Here the character of the deposition and the nature of the sediments fluctuate periodically, with the result that several distinct types of deposits may be recognized in most exposures. The sediments are

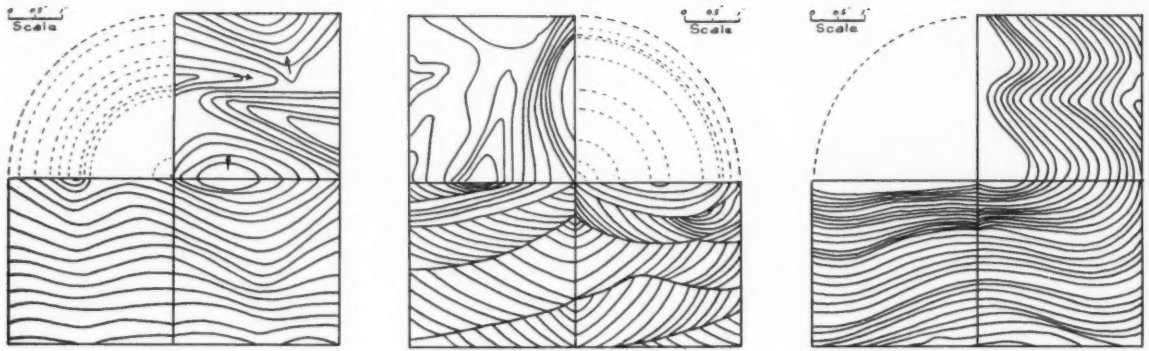


FIG. 4.—Ripple cross lamination pattern in three dimensions. Tracing from typical cases in upper portion of Colorado Delta

dominantly fine sand, but beds and lenses of pure mud are not uncommon. In high banks along the present river near Pescadero Dam the sand layers average about 40 inches and are separated by layers of mud approximately 6 inches thick (see Pl. I, *E*). Fifteen miles farther upstream, near San Luis, the proportion of mud to sand is much

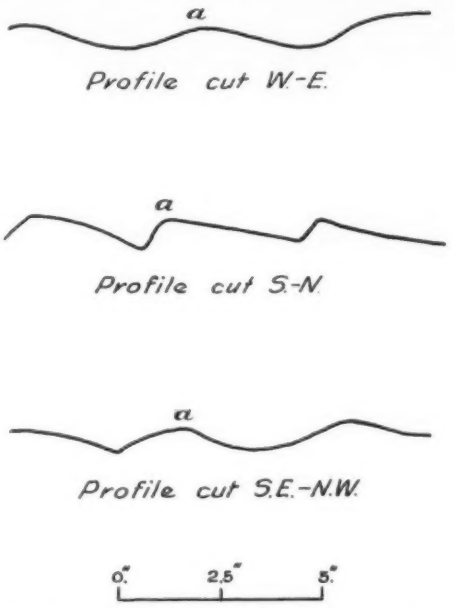


FIG. 5.—Profiles of a rippled surface along three sections cut through the common point (*a*).

less, and deposits of mud more than 2 inches thick have not been observed.

Mud layers in the middle portion of the Colorado River Delta are flat bedded except where deposited over irregular surfaces, and presumably are due to settling under quiet conditions. The interbedded sand layers, on the other hand, are everywhere much cross laminated, having been formed no doubt during times of flood when current action was vigorous and the finer sediments were being carried past into lower parts of the delta. Cross lamination in these

deposits is not only prevalent but also distinctive in character. So far as observed, it consists of two principal types: one is characterized by long, nearly straight laminations dipping at low angles (see Pl. I, *F*); the other is the same type of ripple cross lamination that dominates the original structure of the upper portion of the delta.

Inasmuch as the mud deposits of the middle portion of the delta are either flat bedded or without bedding, their structure needs no description. Ripple cross lamination in the fine-sand deposits has already been described in connection with the upper delta sediments, so the only structures requiring analysis here are the long, sloping laminations of the sand deposits. These represent a smooth traction phase and may best be termed "local delta foreset laminations," as they are formed on the advancing fronts of local deposits. Typical examples near Pescadero Dam average 10 feet in length and dip about 10°, but some observed near San Luis average only 6 feet in length. In all cases the beds dip downstream or in a general southerly direction. Apparently, their origin is comparable to that of foreset beds in fan-shaped or lobate experimental deltas of the type described by Smith,[16] and they are developed as parts of local deltas built up during flood stages. Thus, deposits which belong to the topset beds of the Colorado River Delta actually consist in many places of typical sloping, foreset laminations.

BEDDING IN LOWER PORTION OF SUBAERIAL DELTA

The bedding of the mud deposits in the lower portion of the subaerial delta—that part, including the estuary, south of the margins of recent deltaic fans—is remarkably uniform over a wide area. As seen along the banks of the river for many miles, it consists of very thin, flat-lying laminae, individual layers for the most part averaging less than an inch in thickness. Here texture is as monotonous as structure, and both features clearly indicate the results of slow settling under flood-plain conditions.

Throughout the area are notable exceptions to the usual uniform type of flat bedding. At varying intervals structures have been developed which, in vertical sections normal to the strike, appear as a series of downcurving, concentric arcs. This pattern results from a

[16] *Op. cit.*, pp. 729–42.

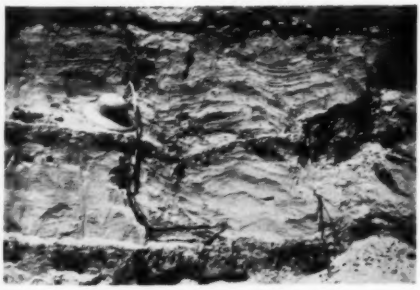
PLATE I

A

B

C

D



E

F

Upper and middle portions of Colorado Delta: *A*, *B*, typical ripple-covered surfaces of silt; *C*, ripple cross lamination, combination of types; *D*, pseudobedding planes due to lee-side concentration in ripple cross lamination; *E*, cycles in mud-silt deposition; *F*, local delta foreset beds.

PLATE II

A

B

C

D

E

F

Lower portion of Colorado Delta showing stages in scour-and-fill deposit of mud: *A*, normal flat-lying laminae; *B*, channel cut through; *C*, beginning of redeposition; *D*, partially refilled channel; *E*, completely refilled channel; *F*, pseudobedding developed in mud deposits.

group of layers which form shells of a roughly semicylindrical shape. In general appearance they are analogous to beds in a syncline, but from point of view of origin they are distinctly different. They are developed when sloughs or small channels are cut back through the banks of the river by the runoff of tide waters and are subsequently filled with mud layers which conform to the shape of the channel. The method of developing this type of bedding is illustrated in Plate II, *A–E*, which shows stages in erosion and redeposition.

The concentric arc pattern commonly displayed along riverbanks is a cross section of the structure described above. Cut at lower angles, these deposits will show a different pattern, the width of the arcs increasing until, in sections cut longitudinally, the beds are essentially parallel and, because of the very slight gradient of the original channel in most cases, almost horizontal.

The average bed of mud in most of the lower portion of the subaerial delta is 1 or 2 inches thick, but in some localities structureless layers many feet in thickness are found. The cause of the parting planes appears to be related to pauses in deposition with accompanying concentration of salts which form a crust on each surface. Furthermore, although only in a very few instances can any textural differences be noted between beds, it is probable that each break or pause in the deposition of this material means a change in grain size, and, because of the extreme fineness of all detrital matter in this area, such differences are difficult to detect. The time represented by the individual beds or layers has not yet been determined by the writer.

Vegetation is scarce over most of the lower portion of the Colorado River Delta because of the concentration of salines. Locally, however, it has been responsible for partially destroying original structures and at the same time developing a type of "mud tube." In some instances, while sending down roots, it has caused the mud layers to bend sharply downward in areas immediately surrounding them. In many places these roots are seen to have been subsequently destroyed and the resulting cavities to have been filled in by mud from above, thus developing a structure which in cross section ap-

pears as a vertical mud column dissecting the normal laminated structure.

A type of pseudobedding occurs locally in mud deposits of the Colorado River Delta. It consists of irregular shells that develop around lumps or masses of the same material and in vertical section usually has the appearance of highly folded beds (Pl. II, *F*). Detailed studies of this type of structure have not been made, but field observations indicate clearly that it is a feature due to shrinkage following the drainage of water from the surface of the area. No textural or compositional variations appear to be involved in its development.

STATUS AND VALUE OF STUDY

In this paper the writer makes no pretense of giving a complete survey of the original structures of deltas or even of setting forth all those which comprise this particular delta. Because, however, information which may be important in understanding certain geological formations has been obtained and because the study may be extended indefinitely before complete data are obtained, it is considered desirable to publish at this time the findings of the first seasons of field investigation.

There are several aspects of the Colorado Delta structures concerning which it is hoped knowledge can be obtained by future study. First, it is desirable that information be gathered on the time represented by the beds and laminations in various parts of the delta. Second, detailed studies should be made of structures in the border areas of the delta, especially where wind deposits mingle with those of the river and where the beach and flood plains merge. Lastly, data should be obtained on lamination of the foreset portions of the delta for comparison with that in the topset beds.

Although bedding in the subaerial portions of the Colorado Delta may not be, and probably is not, typical of that in corresponding parts of all great deltas, nevertheless certain aspects of the data presented in this paper should be of value in interpreting subaerial environments and in recognizing deposits of former deltas. It is significant that the uppermost portion of the subaerial delta studied is characterized by nearly pure sand in contrast to the lower portion

which is of nearly pure mud; that the mingling of these types toward the delta center involves no very apparent gradation in clastic grain size or mixing of types, but rather a conspicuous alternation of beds of each type; that the relative thickness of mud beds and sand beds in given sections reflects the proximity to one or the other extremes of the delta.

Another important result of the present delta studies is the establishing of certain types of cross lamination as characteristic of subaerial deposition on large deltas of the Colorado type. Especially diagnostic are the very widespread and intricate ripple cross lamination patterns which are found throughout a considerable portion of all sand deposits over many parts of the delta. Likewise constituting a large percentage of all deposits and difficult to duplicate in any other environment are the remarkably uniform, flat-bedded mud layers with scour-and-fill channels of semicylindrical shape developed through them at intervals.

Perhaps the greatest value of the present study is the fact that it represents the beginning of systematic recording of original structures in a large delta. Only after a considerable amount of such work on many deltas and on other parts of deltas, can a clear understanding be obtained of the meaning of the many and diverse types of bedding found in various geological formations which represent deltaic deposits. Inasmuch as these constitute an appreciable percentage of all deposits in the geologic record, it is obvious that this is a fruitful course of study to pursue.

ACKNOWLEDGMENTS.—Acknowledgment is made to the Carnegie Institution and especially to its president, Dr. John C. Merriam, for making possible this investigation through a research grant. For many helpful suggestions, thanks are given Mr. F. E. Matthes, of the U.S. Geological Survey, and to Professors C. M. Nevin and H. Ries, of Cornell University. The drawings are by Louis Shellbach.

SOME METAMORPHIC PHENOMENA PRODUCED BY GABBROIC INTRUSION

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ABSTRACT

Locally, metamorphism by the Duluth gabbro has caused progressive changes in texture and mineralogy of the Knife Lake slate, producing a granulitic and granoblastic rock very similar to the gabbro in composition. Several metamorphic zones may be roughly delineated. These zones are characterized by mineral suites different from those developed in zones surrounding granite and especially by the rather abundant development of orthopyroxene in some zones. This mineral development is attributed to the influence of the gabbroic magma and is thought to represent a characteristic type of metamorphism produced in argillaceous rocks by such magma.

INTRODUCTION

Characteristic minerals produced by the metamorphism of argillaceous rocks have been used to establish metamorphic facies and zones,[1] and the importance of progressive metamorphic changes has been emphasized recently in the excellent papers by Barth[2] and by Billings.[3] In the areas investigated by Barth and by Billings, staurolite, cyanite, and sillimanite are important zonal minerals. In each case there is a progressive increase in intensity of metamorphism toward the southeast, which is attributed to igneous intrusion, the dominant type of intrusive being granite.

Several years ago Grout called attention to some important differences in metamorphism produced by gabbro and by granite intrusions and described, as a typical example, the metamorphism produced in slates of Minnesota.[4] It is worthy of note that staurolite, cyanite, and sillimanite are almost wholly absent from the mineral

[1] P. Eskola, "The Mineral Facies of Rocks," *Norsk. Tidsskv.*, Vol. VI (1920), p. 146; C. E. Tilley, "The Facies Classification of Metamorphic Rocks," *Geol. Mag.*, Vol. LXI (1924), pp. 167–71.

[2] Tom F. W. Barth, "Structural and Petrologic Studies in Dutchess County, N.Y.," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), pp. 777–79.

[3] Marland P. Billings, "Regional Metamorphism of the Littleton–Moosilauke Area, New Hampshire," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 470–75, 539–44.

[4] F. F. Grout, "Contact Metamorphism of the Slates of Minnesota by Granite and by Gabbro Magmas," *Bull. Geol. Soc. Amer.*, Vol. XLIV (1933), pp. 989–1040.

suites described by him. This is significant, and it is the writer's belief that this difference should be emphasized.

During the summer of 1926 it was the writer's privilege to visit parts of Minnesota under the guidance of the late U. S. Grant and of J. T. Stark. Opportunity was afforded for a study of some of the contact metamorphism produced by the Duluth gabbro, and at this time the writer became particularly interested in certain aspects of the metamorphism produced in the Knife Lake slates. This metamorphism was studied especially in the vicinity of Kekequabic Lake and less extensively around Gabbemichigamma Lake, about ten miles farther east. Specimens were collected and later studied in the laboratory, twenty-three thin sections being cut from selected ones. The results of this study have not been submitted for publication prior to this time because of the incompleteness of the investigation and some question regarding (1) the significance of a brecciated or conglomeratic zone within the Knife Lake slate and (2) the metamorphic effect produced by the Kekequabic granite. It is now believed that the intrusion of the Kekequabic granite presents no difficulty, as it preceded the gabbro intrusion, and produced different and inferior contact effects which were largely obliterated by the later intrusion. Further, it is thought that the significance of the brecciated or conglomeratic zone within the Knife Lake slate does not enter into the present problem, as there is continuity of the same or similar slate until the gabbro is reached.

GENERAL GEOLOGY AND FIELD RELATIONS

Various pre-Cambrian rocks are exposed in the vicinity of Kekequabic and Gabbemichigamma lakes, the principal ones being Ely greenstone, Ogishke conglomerate, Knife Lake slate, Kekequabic granite, Biwabik iron formation, and Duluth gabbro. Of these units the present discussion is concerned chiefly with the Knife Lake slate and the Duluth gabbro, and to a minor extent with the Kekequabic granite. Hence the field relations of these rocks only need be considered.

South of Kekequabic Lake and around Shoofly Lake the Knife Lake slate is intruded by both the Kekequabic granite and the Duluth gabbro, the latter being the younger intrusion (Fig. 1). At

Gabbemichigamma Lake the slate crops out along the northwestern, northern, and northeastern shores, and the gabbro forms the east-central shores, as well as islands, and is in contact with the slate east of the northern part of the lake.

NORMAL KNIFE LAKE SLATE SERIES

The Knife Lake slate series, unmodified by contact action, has been well described by Grout.[5] In general, the rocks are gray to black slates and graywackes composed chiefly of various feldspars and quartz embedded in much fine chloritic material. At Keke-

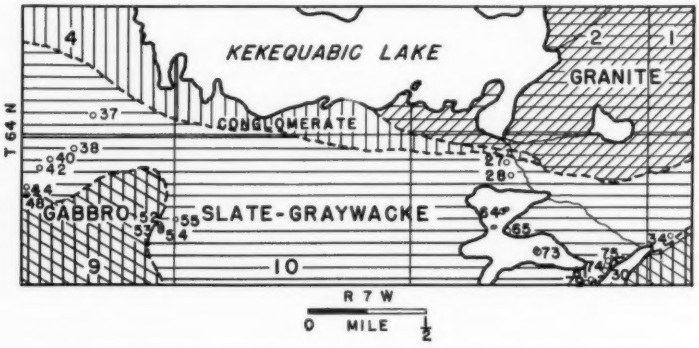


FIG. 1.—Map showing general geology south of Kekequabic Lake and location of specimens listed in Tables 1, 2, and 3. Southeast of Kekequabic Lake are River Lake and Shoofly Lake. Geology after *U.S. Geol. Surv. Mono. 45*, slightly modified.

quabic Lake the Knife Lake series has been separated into three types that grade into each other and in the field have been designated slate-graywacke, agglomerate, and green schist.[6] The agglomeratic and green-schist types contain large amounts of rather fibrous, colorless to light-green amphibole. Although outcropping south of Kekequabic Lake, these latter phases apparently do not occur within the metamorphic zones considered.

[5] *Ibid.*, pp. 992–95.

[6] U. S. Grant, "The Geology of Kekequabic Lake in Northeastern Minnesota with Special Reference to an Augite Soda-Granite," *Geol. and Nat. Hist. Surv. Minn. 21st Ann. Rept.* (1892), pp. 19–27.

METAMORPHISM

MEGASCOPIC CHARACTERISTICS

In the field a definite contact zone of a somewhat gradational nature is apparent around the Duluth gabbro south of Kekequabic Lake and east of Gabbemichigamma Lake. This contact zone at Kekequabic Lake is shown on the early geologic map of the area.[7] Conditions vary somewhat within this zone, as the gabbro, which forms a lopolith, once overlay the slate in certain places. Subsequently it was eroded, and the metamorphism of the slate in those places was caused by the overlying gabbro rather than by the gabbro now exposed much farther away horizontally.[8]

In general, early stages of metamorphism are here characterized by the appearance of small veins, by slight textural changes, and by incipient mineral formation, but with increasing metamorphism both textural changes and mineral development become prominent features. At a distance of about 3,300 feet from the gabbro contact southwest of Kekequabic Lake, and about 3,000 feet at Gabbemichigamma Lake, the Knife Lake slate is cut by an intricate network of very small veins composed chiefly of quartz. It is thought that this intricate veining represents the first evidence of the proximity of the gabbro. On approaching the gabbro after passing this horizon, the sediments gradually lose their normal aspect. They become granular and assume a brownish cast and a peculiar saccharoidal appearance, a characteristic which early caused them to be called "muscovado."[9] Metamorphic effects become quite apparent at distances varying between 900 and 750 feet from the contact.

The area which best shows gradual progressive metamorphism is southwest of Kekequabic Lake (Fig. 1). In that area, at a distance of about 1,000 feet from the gabbro, small flakes of brown biotite are apparent if the rock is examined with a lens. With closer approach to the contact these biotite flakes are larger and are visible with the

[7] J. M. Clements, "The Vermilion Iron-bearing District of Minnesota," *U.S. Geol. Surv. Mono. 45* (1903), Atlas, geology sheet 16.

[8] Grout, "The Lopolith: An Igneous Form Exemplified by the Duluth Gabbro," *Amer. Jour. Sci.*, Vol. XLVI (1918), pp. 516–22; "Contact Metamorphism . . . . ," *op. cit.*, pp. 1008–9.

[9] A. N. Winchell, "Report of a Geological Survey in Minnesota during the Season of 1886," *Geol. and Nat. Hist. Surv. Minn. 15th Ann. Rept.* (1886), p. 183.

unaided eye, and a lens discloses the presence of small feldspar crystals. Within 50 or 75 feet of the contact, which is somewhat gradational, the mica, in some instances, is a quarter of an inch across. Nearer the contact, however, mica is not visible, and the rock is penetrated by numerous small dikelets from the gabbro. The dikelets vary in width from a fraction of an inch to a few inches and are pegmatitic in character, although of much the same composition as the gabbro. Some of them, however, contain quartz and hornblende, with some pyroxene, as major constituents, and plagioclase in minor amounts only.

In the vicinity of Gabbemichigamma Lake conditions are much the same, but to the southeast of Kekequabic Lake, around River Lake, they are much more variable. In the latter vicinity, at distances as much as one-half mile from the contact, the rocks appear to be much more highly metamorphosed than at a similar distance from the contact southwest of Kekequabic Lake. Moreover, on small islands in the southeastern part of River Lake the character of the metamorphosed slate is about the same as that observed within 50 feet of the contact farther west. These facts indicate that at the time of metamorphism there must have been gabbro in this vicinity much closer than the gabbro boundary now visible.

TEXTURAL CHANGES

Thin sections show well the changes in texture produced by contact metamorphism. The clastic texture of the normal slate passes into granulitic, granoblastic, and poikiloblastic textures, with gradual increase in grain size (Pl. I, *A–D*). Specimens from the gabbro-slate contact exhibit a composite texture, one part being granulitic and the other part showing the texture of the normal gabbro with its elongated plagioclase (Pl. I, *E*); it is therefore possible to distinguish metamorphosed sediment from gabbro on the basis of texture in thin section, a fact observed by Grout, also.[10] Some specimens from near the contact show tiny dikelets in thin section (Pl. I, *F*).

It is of interest to note that in the vicinity of River Lake the specimens which appear to represent a higher degree of metamorphism than would be expected if judged by the position of the

[10] "Contact Metamorphism . . . . ," *op. cit.*, pp. 1011, 1012, 1015, and 1032.

visible contact, show textures comparable to specimens collected near the visible gabbro-slate boundary southwest of Kekequabic Lake and around Shoofly Lake.

MINERALOGICAL CHANGES

Petrographic study reveals that mineralogical changes parallel, in a general way, the textural changes noted. The chief minerals produced by contact metamorphism of the slate in this area are plagioclase (andesine to labradorite), hypersthene or bronzite, augite, biotite, and hornblende. The metamorphosed slates contain, also, magnetite, tourmaline, diopside, hedenbergite, and staurolite in subordinate amounts; accessory quartz, apatite, zircon, pyrite, and pyrrhotite; and some secondary muscovite. Not all these minerals occur in a single specimen, and staurolite was found in but one thin section. It is believed that the principal minerals may be grouped roughly into zones indicative of different degrees of metamorphism. Both from field observation and from microscopic study the area which apparently furnishes the best evidence of mineral zoning is the one southwest of Kekequabic Lake. It shows, also, the best textural gradation. The mineral development in the slate of this area will be discussed first, and the changes produced elsewhere will be compared with it.

The approximate percentages of the chief minerals noted in thin sections from specimens southwest of Kekequabic Lake are shown in Table 1. The location of specimens is shown in Figure 1. From the tabulation it will be noticed that the rocks may be placed roughly in three groups, each characterized by a definite mineral assemblage. These groups are as follows:

1. Hornblende-biotite-plagioclase group, 1,450–650 feet from gabbro
2. Hypersthene-biotite-plagioclase group, 650–100 feet from gabbro
3. Hypersthene-augite-plagioclase group, 70–0 feet from gabbro

Tourmaline was noted in but two of these specimens, Nos. 38 and 40, at distances of 900 feet or more from the gabbro.

A few details of the minerals in these groups may be of interest. The plagioclase, newly crystallized, varies between andesine and labradorite, rarely becoming as calcic as the latter. It is more abundant near the gabbro, and there is some indication that it tends to be

more calcic. In some specimens both hypersthene and biotite are poikiloblastically developed, the former around plagiocalse, the latter around plagioclase in some specimens and around hypersthene in others. Poikiloblastic development of hypersthene was not observed in specimens 70 feet or less from the contact, or in all specimens farther away. If not poikiloblastically developed, hypersthene occurs as droplike and rounded crystals. Augite showed no poikiloblastic development in the specimens studied. Hornblende occurs in

TABLE 1

APPROXIMATE MINERAL COMPOSITION OF SPECIMENS SOUTHWEST OF KEKEQUABIC LAKE

No.	Distance from Contact (in Feet)	Hornblende (%)	Pyroxene		Plagioclase (%)	Biotite (%)	Magnetite (%)	Quartz (%)	Orthoclase (%)
			Orthorhombic (%)	Monoclinic (%)					
38	1,450–900	40	.....	.....	.....	15	tr.	40	.....
40	900	.....	.....	.....	15	50	.....	25	10
42	650	5	35	.....	15	5	10	30	.....
44	170	.....	5	.....	50	40	.....	.....	5
55	100	.....	50	.....	35	10	tr.	5	.....
54	70	.....	30	30	38	tr.	2	.....	.....
53	55	.....	25	20	50	tr.	5	.....	.....
52	40	.....	50	.....	50	.....	.....	.....	.....
48	0	.....	30	10	58	2	tr.	.....	.....

droplike crystals in this area but has a different development elsewhere.

In five specimens from Gabbemichigamma Lake, at distances of 200, 300, 700, and 750 feet from the visible gabbro contact, mineral development is much the same as in the area southwest of Kekequabic Lake. However, no hypersthene was noted, the pyroxene being chiefly augite and diopside, with some hedenbergite. Also, tourmaline is contained in three specimens collected at distances of 700 and 750 feet from the contact, but not in two specimens within 200 and 300 feet of the gabbro. Tourmaline is much more abundant in these specimens than in those collected around Kekequabic Lake, and in one instance could be distinguished with a hand lens. The fact that the gabbro shows pegmatitic development around Gab-

bemichigamma Lake is of interest in this connection. Of interest, also, is the lack of tourmaline near the gabbro and its development at a distance, a fact also observed by Grout.[11]

Southeast of Kekequabic Lake, around River Lake and Shoofly Lake, mineral distribution with respect to distance from the visible gabbro contact does not show the uniformity observed southwest of Kekequabic Lake. The approximate mineral percentages in speci-

TABLE 2

APPROXIMATE MINERAL COMPOSITION OF SPECIMENS SOUTHEAST OF KEKEQUABIC LAKE

No.	Distance from Contact (in Feet)	Hornblende (%)	Pyroxene		Plagioclase (%)	Biotite (%)	Magnetite (%)	Quartz (%)	Orthoclase (%)
			Orthorhombic (%)	Monoclinic (%)					
27	3,290	50	20	.....	5	5	tr.	10	10
28	3,065	50	5	5	10	10	.....	10	10
63	2,640	30	35	5	10	5	tr.	5	10
64	2,640	50	.....	5	15	10	5	5	10
65	2,375	15	50	.....	25	5	2	.....	.....
73	1,450	.....	.....	.....	60	15	10	2	10
34	790	35	20	15	20	.....	.....	5	.....
30*	260	.....	10	.....	60	10	10	.....	.....
74	50–0	.....	25	20	50	2	3	.....	.....
79	50–0	.....	30	.....	50	tr.	5	10	.....
81	50–0	.....	30	10	50	5	5	.....	.....

* This specimen contains about 10 per cent of staurolite.

mens from this vicinity are shown in Table 2, and the location of specimens is shown in Figure 1. Certain similarities to Table 1 are at once apparent—hornblende absent very near the contact; biotite persisting slightly nearer the contact than hornblende, but also being practically absent within 50 feet of the gabbro; hypersthene and augite present together, for the most part, only near the contact; and increase in amount of plagioclase close to the gabbro. The chief points of difference are (1) the occurrence of hypersthene with hornblende, both in relatively large amounts; (2) the great distance from the visible contact at which both hypersthene and plagioclase occur; and (3) the lack of regular gradation toward the present gabbro

[11] *Ibid.*, p. 1013.

boundary, except in a general manner. How much of this apparent irregularity is due to the fact that the progressive mineral sequence observed southwest of Kekequabic Lake may not represent the true gradation toward the actual metamorphic contact is not known.

TABLE 3

COMBINATION OF TABLES 1 AND 2

No.	Distance from Contact (in Feet)	Hornblende (%)	Pyroxene		Plagioclase (%)	Biotite (%)	Magnetite (%)	Quartz (%)	Orthoclase (%)
			Orthorhombic (%)	Monoclinic (%)					
27	3,290	50	20	.....	5	5	tr.	10	10
28	3,065	50	5	5	10	10	.....	10	10
63	2,640	30	35	5	10	5	tr.	5	10
64	2,640	50	.....	5	15	10	5	5	10
65	2,375	15	50	.....	25	5	2	.....	.....
73	1,450	.....	.....	.....	60	15	10	2	10
38	1,450–900	40	.....	.....	.....	15	tr.	40	.....
40	900	.....	.....	.....	15	50	.....	25	10
34	790	35	20	15	20	.....	.....	5	.....
42	650	5	35	.....	15	5	10	30	.....
30*	260	.....	10	.....	60	10	10	.....	.....
44	170	.....	5	.....	50	40	.....	.....	5
55	100	.....	50	.....	35	10	tr.	5	.....
54	70	.....	30	30	38	tr.	2	.....	.....
53	55	.....	25	20	50	tr.	5	.....	.....
52	40	.....	50	.....	50	.....	.....	.....	.....
74	50–0	.....	25	20	50	2	3	.....	.....
79	50–0	.....	30	.....	50	tr.	5	10	.....
81	0	.....	30	10	50	5	5	.....	.....
48	0	.....	30	10	58	2	tr.	.....	.....

* This specimen contains about 10 per cent of staurolite.

However, conditions are much more uniform in that area, so it seems very likely that the discrepancy in this southeastern area is due to the fact that the visible contact was not the nearest contact at the time metamorphism was produced. Moreover, the texture of some of these specimens and the occurrence of small dikelets in the outcrops indicate that their position in the metamorphic sequence is much nearer the gabbro than is shown by the visible boundary.

In the River Lake–Shoofly Lake area, hornblende, hypersthene, and biotite are poikiloblastically developed, usually around plagioclase. This development was not noted in augite.

MINERAL ZONES

The restricted nature of the present investigation does not permit definite conclusions regarding the establishment of mineral zones indicative of varying degrees of contact metamorphism. Nevertheless, certain occurrences are highly suggestive and may furnish an indication of the type of progressive metamorphism to be expected around the larger gabbro masses. A combination of Tables 1 and 2 (Table 3) suggests, despite some irregularities, a zoning for this particular area. If the mineral development in the area southwest of Kekequabic Lake be considered as the most normal progressive sequence, which appears to be warranted by the progressive changes in texture and by the development of gabbroic dikelets in the metamorphosed rocks, a tentative zonal arrangement based on mineral constitution, texture, and presence of small dikelets is suggested. This zoning appears to be as shown in Table 4, the zones being ar-

TABLE 4

Zone	Width (in Feet)	Characteristic Minerals				
1	50–100	Plagioclase	.....	.....	Hypersthene	Augite
2	500–600	Plagioclase	Biotite	.....	Hypersthene	.....
3	200–300	Plagioclase	Biotite	Hornblende	Hypersthene	.....
4	400–500	Plagioclase	Biotite	Hornblende	.....	.....

ranged in order of increasing distance from the gabbro. The plagioclase in these zones is between andesine and labradorite in composition. Magnetite is present in all zones, but perhaps is more abundant in the zones close to the gabbro. These estimates would fix the width of the more evident contact zone between 1,150 and 1,500 feet.

CHEMICAL CHANGES

No chemical analyses of specimens studied were obtained. It seems apparent, however, from the mineralogical changes, that there must have been introduction of magnesia and lime in some amount, and probably introduction of iron oxide as well. These and other chemical changes were shown, by Grout's analyses,[12] to have taken place in Minnesota slates.

MINERALOGICAL COMPARISONS

Many descriptions of metamorphosed argillaceous rocks emphasize biotite, garnet, staurolite, andalusite, cordierite, cyanite, and

[12] *Ibid.*, p. 1017.

sillimanite as typical minerals, these minerals being accompanied by more or less feldspar, especially in the intermediate and high-grade metamorphic zones. For instance, Barth[13] cited the sequence biotite, garnet, staurolite, cyanite, sillimanite as indicative of increasing degrees of metamorphism; and Billings[14] observed, with progressive metamorphism, the formation of biotite, garnet, staurolite, and sillimanite, but noted also the development of some andalusite near a quartz-monzonite intrusion. Many other writers have shown that some or all of these minerals are characteristic of metamorphosed argillaceous formations. This is well known. However, in those cases in which an intrusion is mentioned, it is practically always a silicic one, granite being most common. The more basic intrusions, such as gabbro, apparently give rise to a different mineral development in many, if not in most, instances.

Minerals which may be cited as characteristic of argillaceous formations metamorphosed by gabbroic intrusions include biotite, cordierite, orthopyroxene (hypersthene or bronzite), some monoclinic pyroxene (augite, diopside, hedenbergite), hornblende, and plagioclase near labradorite in composition. Various investigators[15] have noted some or all of these minerals in the metamorphosed slates around the Duluth gabbro. Orthopyroxene, because of its comparative rarity as a metamorphic mineral, is perhaps the most distinctive mineral of this group.

Orthopyroxene has been stated to form in various ways. Harker[16] states that it may be formed by "thermal" metamorphism of argillaceous rocks rich in chlorite and containing calcareous material, or, if formed under high-grade conditions, by the further metamorphism

[13] *Op. cit.*, pp. 777–79. [14] *Op. cit.*, pp. 539–44.

[15] W. S. Bayley, "The Eruptive and Sedimentary Rocks of Pigeon Point, Minnesota, and Their Contact Phenomena," *U.S. Geol. Surv. Bull. 109* (1893), p. 100; Winchell, "Mineralogical and Petrographic Study of the Gabbroid Rocks of Minnesota," *Amer. Geologist*, Vol. XXVI (1900), pp. 294–306; Grant, "Contact Metamorphism of a Basic Igneous Rock," *Bull. Geol. Soc. Amer.*, Vol. XI (1900), pp. 503–10; A. H. Elftman, "The Geology of the Keweenawan Area in Northern Minnesota," *Amer. Geologist*, Vol. XXII (1898), pp. 147–49; Clements, *op. cit.*, p. 419; M. L. Nebel, "The Basal Phases of the Duluth Gabbro near Gabamichigami Lake, Minnesota, and Its Contact Effects," *Econ. Geol.*, Vol. XIV (1919), pp. 394–95; Grout, "Contact Metamorphism . . . . ," *op. cit.*

[16] Alfred Harker, *Metamorphism* (London: Methuen & Co., Ltd., 1932), pp. 53, 57.

of biotite, the potassic part of this mineral, plus some aluminous silicate, yielding orthoclase, and hypersthene being developed from the ferromagnesian constituents. Again, its development near Kristiania apparently was dependent on gradation from shale to marl.[17] However, in other instances orthopyroxene has been formed because of the character of the intrusive magma. Brögger[18] observed that hypersthene was developed in the Ogygia shales near the contact of a basic plutonic rock (olivine-gabbro-diabase) but that the same shales metamorphosed by nordmarkite contained no hypersthene; he concluded that the difference was due to the character of the intrusives, indicating that magnesia and iron oxides were added from the basic magma. Grout[19] has shown that the percentages of lime, magnesia, iron oxides, alumina, and titania are greater in metamorphosed slates of Minnesota near the Duluth gabbro than in the original formations and attributes these increases to additions from the gabbro magma. The increase of magnesia and iron oxide undoubtedly accounts, at least in part, for the formation of the orthopyroxene, and the increase in lime would contribute to the formation of basic plagioclase. Bowen[20] has shown that, if argillaceous inclusions are immersed in basaltic magma, the formation of orthopyroxene and basic plagioclase should result, and Read[21] has demonstrated that aluminous inclusions in basaltic magma give rise to the formation of norite and that an intermediate stage in the norite development is the production of xenoliths composed chiefly of hypersthene and basic plagioclase. Others, also, have postulated that norite may be formed by assimilation of argillaceous material

[17] V. M. Goldschmidt, *Die Kontaktmetamorphose im Kristianiagebeit* (Videnskapsselskapets Skrifter I. Mat.-Naturv. Klasse, No. 1 [1911]), p. 140. Consult G. W. Tyrrell, *The Principles of Petrology* (London: Methuen & Co., Ltd., 1926), p. 295, for a summary statement and tabulation.

[18] W. C. Brögger, "The Basic Eruptive Rocks of Gran," *Quar. Jour. Geol. Soc.*, Vol. L (1894), p. 21.

[19] "Contact Metamorphism . . . . ," *op. cit.*, pp. 1018–19, 1022–23, and 1032.

[20] N. L. Bowen, *The Evolution of the Igneous Rocks* (Princeton: Princeton University Press, 1928), pp. 207–16.

[21] H. H. Read, "The Gabbros and Associated Xenolithic Complexes of the Haddo House District, Aberdeenshire," *Quar. Jour. Geol. Soc.*, Vol. XCI (1935), pp. 609, 620–21, 624–30, and 634.

by basic magmas. Moreover, Grout[22] has recently emphasized the fact that slate inclusions in the Duluth gabbro have been converted into a rock resembling a gabbro in composition.

In addition to these investigations, attention should be called to some early descriptions of certain peripheral and contact phases of the Duluth gabbro. Bayley[23] described various peripheral phases of the gabbro and cited several varieties: (1) olivinitic; (2) hypersthenitic; (3) biotitic and hornblendic. The hypersthenitic variety contains chiefly hypersthene, diallage, and plagioclase similar in composition to that occurring in the main gabbro mass. This variety contains, also, some biotite and hornblende. In the biotitic and hornblendic varieties the pyroxenes decrease in amount, and there is a corresponding increase in the amount of biotite and hornblende. The excellent descriptions of these varieties show that some of the pyroxene, hornblende, and biotite has a cellular development and includes other minerals, this development being very similar to that described for certain contact rocks, a fact recognized by Bayley. It is worthy of note that these rocks were described as *peripheral* phases of the gabbro and that some other investigators thought they were sediments changed by contact metamorphism. In either case, the mineral development is distinctive and is probably a contact phenomenon, produced either by assimilation or by contact metamorphism.

It was early recognized that in some instances, at least, distinctive minerals were produced by intrusion of the Duluth gabbro into the surrounding rocks.[24] One of the important facts presented is that hypersthene is characteristic of metamorphism by gabbro but not by granite.

It seems clear from the foregoing discussion that orthopyroxene

[22] "Criteria of Origin of Inclusions in Plutonic Rocks," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 1556–58.

[23] "The Basic Massive Rocks of the Lake Superior Region," *Jour. Geol.*, Vol. III (1895), pp. 1–20.

[24] Bayley, "The Eruptive and Sedimentary Rocks . . . . ," *op. cit.*, pp. 100–104; Winchell, "Structural and Petrographic Geology of the Taconic and Archean," *Geol. and Nat. Hist. Surv. Minn. Final Rept.*, Vol. V (1900), p. 11; Grant, "Contact Metamorphism . . . . ," *op. cit.;* C. K. Leith, "The Mesabi Iron-bearing District of Minnesota," *U.S. Geol. Surv. Mono. 43* (1903), p. 164.

and basic plagioclase may be characteristic of argillaceous rocks metamorphosed by gabbro, whereas staurolite, cyanite,[25] sillimanite, and andalusite may be typical of metamorphism produced by granite. There appears to be no reason why these latter minerals should not form around a gabbro intrusion if it were emplaced under the same conditions as a granite, unless it be due to chemical environment and addition of material. Indeed, staurolite and andalusite (?) have been noted in minor amounts in the metamorphosed slates around the Duluth gabbro by the writer and others. Also, staurolite, sillimanite, and cyanite have been noted by Williams[26] and by Rogers[27] in the metamorphic rocks produced by the intrusion of the "Cortlandt series" into the Manhattan schist. However, at the contact these minerals disappear, for the most part, and the rock is composed chiefly of feldspar, colorless pyroxene, and garnet.[28] Bowen,[29] discussing the effect of basaltic magma on aluminous inclusions, using this case of intrusion into the Manhattan schist as an example, has shown that after inclusions very rich in sillimanite have been produced, a turning-point is reached, resulting in the formation of orthopyroxene and basic plagioclase. From a consideration of these facts, it would appear that the chemical environment of gabbro magma should promote the formation of hypersthene and plagioclase rather than staurolite, sillimanite, cyanite, and andalusite, at least in the later stages of metamorphism.

It has been suggested by Grout[30] that gabbro magmas may be emplaced under conditions different from those of granite magmas, deformation being largely lacking in the former case, since hornfelses are more characteristic of gabbro metamorphism, and schists more characteristic of granite metamorphism. Hence the mode of emplacement of gabbro might tend to inhibit the production of

[25] Objection may be raised to the inclusion of cyanite as a contact-metamorphic mineral, but such an origin certainly is indicated in some instances.

[26] George H. Williams, "The Contact-Metamorphism Produced in the Adjoining Mica Schists and Limestones by the Massive Rocks of the 'Cortlandt Series' near Peekskill, New York," *Amer. Jour. Sci.*, Vol. XXXVI (3d ser., 1888), pp. 254–69.

[27] G. S. Rogers, "Geology of the Cortlandt Series and Its Emery Deposits," *Ann. N.Y. Acad. Sci.*, Vol. XXI (1911), pp. 11–86.

[28] Williams, *op. cit.*, pp. 255–58.

[29] *Op. cit.*, pp. 207–10.

[30] "Contact Metamorphism . . . . ," *op. cit.*, pp. 1025–26.

minerals such as staurolite and cyanite, whereas the mode of emplacement of granite might tend to promote their formation.

CONCLUSIONS

Intrusion of the Duluth gabbro into the Knife Lake slate around Kekequabic Lake and Gabbemichigamma Lake produced progressive changes in the texture and the minerals of the slate. Granulitic, granoblastic, and poikiloblastic textures were formed, and the grain size increased toward the gabbro, attained a maximum, and decreased slightly at the immediate contact. The chief minerals produced are plagioclase (andesine to labradorite), hypersthene, monoclinic pyroxenes (augite, diopside, hedenbergite), biotite, and hornblende. In the vicinity of Kekequabic Lake a rough metamorphic zoning is indicated. From the gabbro outward the zones are characterized by (1) plagioclase, hypersthene, and monoclinic pyroxene; (2) plagioclase, hypersthene, and biotite; (3) plagioclase, hypersthene, biotite, and hornblende; (4) plagioclase, biotite, and hornblende. Tourmaline, noted in some specimens, was not observed at distances less than 700 feet from the gabbro.

Staurolite, sillimanite, andalusite, and cyanite, characteristic minerals of argillaceous rocks metamorphosed by granite, are almost entirely lacking in the mineral assemblages developed by the Duluth gabbro.

It is thought that a distinctive mineral assemblage characteristic of argillaceous rocks metamorphosed by the larger gabbro masses, and different from the assemblage produced by granite intrusion, is indicated by this and other investigations. The minerals more usually formed by gabbro intrusion include basic plagioclase, orthorhombic pyroxene, monoclinic pyroxenes, biotite, hornblende, and cordierite. Exceptionally, olivine, spinel, and corundum may be produced.[31]

Not all the conclusions stated by the writer are new. It was early recognized by some investigators that basic intrusions produce a different type of metamorphism from acidic ones, and the subject

[31] Grout, "Contact Metamorphism . . . . ," *op. cit.*, pp. 1012–13; Bowen, *op. cit.*, pp. 210–12; Read, *op. cit.*, pp. 624–26.

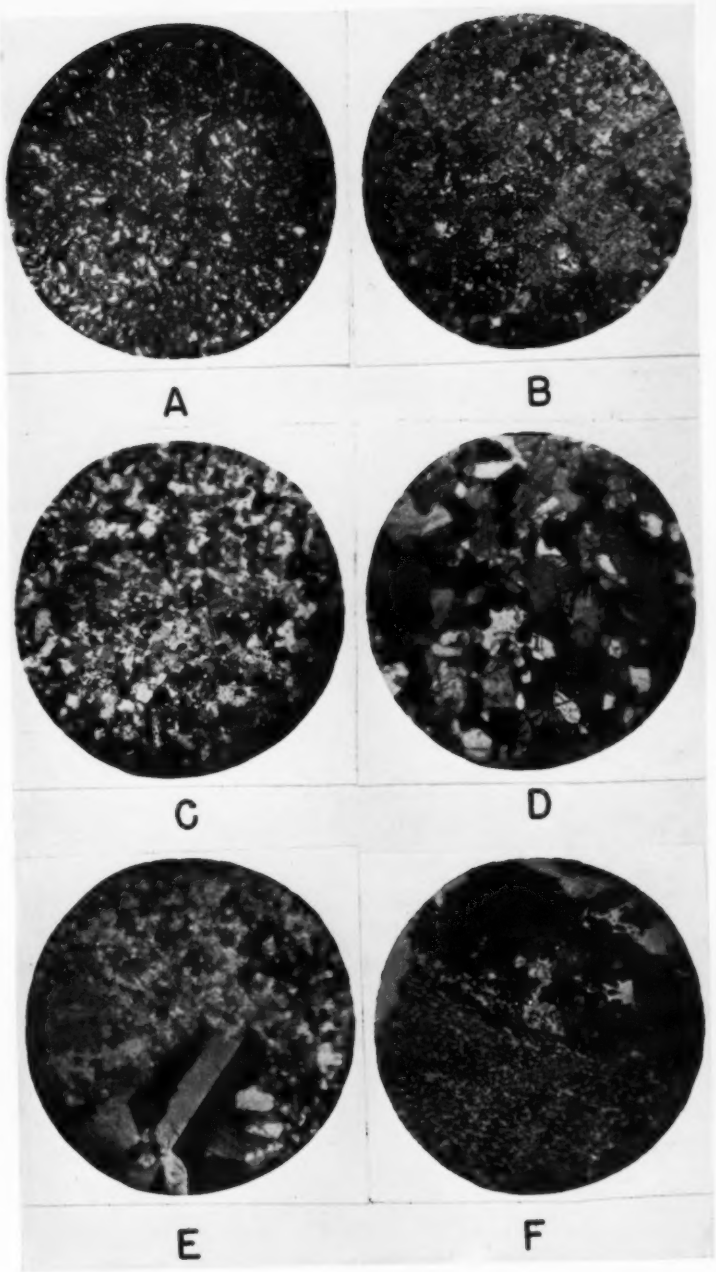
has been reviewed somewhat recently.[32] It is believed, however, that this fact needs emphasis and that more detailed investigations of the contact effects produced by the larger gabbro masses of the world are desirable. The zoning suggested in this paper is but a tentative one, and more detailed investigations from other regions may aid in establishing a definite metamorphic sequence for gabbro intrusions. At the same time, such investigations may further our knowledge of the origin of norite and the conditions under which hypersthene is formed.

[32] Grout, "Contact Metamorphism . . . . ," *op. cit.*, pp. 989–1039.

EXPLANATION OF PLATE I

Photomicrographs showing progressive changes in texture of Knife Lake slate. Crossed nicols × 23 except *F*, which is × 8. *A*, normal slate, 2½ miles from gabbro; *B*, No. 42, 650 feet from gabbro (shows poikiloblastic hypersthene); *C*, No. 30, 260 feet from gabbro; *D*, No. 53, 65 feet from gabbro; *E*, No. 81, at gabbro contact (the coarse part, showing large plagioclase, is gabbro); *F*, No. 52, 40 feet from gabbro (upper part shows dikelet from gabbro, lower part shows metamorphosed slate containing much hypersthene; magnification is lower than other illustrations).

PLATE I



NOMENCLATURE OF GEYSER ERUPTIONS

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ABSTRACT

Records of geyser eruptions show lack of standardization of nomenclature and of technique of observation. The result has been misinterpretation of phenomena. The writer suggests standard terms and technique.

CYCLE

The complete round of geyser activity, from any given point to the recurrence of the same point, is the geyser cycle, which, in turn, is divided into phases.

PREMONITORY PHASE

That portion of the cycle in which it is apparent an eruption is approaching is the premonitory phase. It begins with the first indication that the quiet phase has ended and ends with the appearance of the first spurt of water thrown above the mouth. This phase is divided into stages.

Filling stage.—Begins at the moment it is apparent the water has begun to rise above the level of the quiet phase. Ends at the moment of overflow.

Overflow stage.—The moment when water for the first time overflows the lip.

Flow stage.—That portion of the phase extending from the overflow stage to the beginning of the "preliminary phase."

PRELIMINARY PHASE

That portion of the cycle during which water is discharged from time to time above the geyser mouth, or, in certain cases, puffs of steam are emitted, is regarded as the preliminary phase. It begins with the appearance of the first spurt of water or puff of steam above the mouth.

The phase is divided into the proper number of preliminary spurts of water (or puffs, if steam), each followed by its preliminary pause.

Each set is serially numbered. A spurt is timed from the appearance of the water above the mouth until its disappearance below the mouth. A puff is timed from its appearance above the mouth until the mouth ceases to emit the steam as a definite unit. A pause is timed from the end of either a spurt or a puff to the appearance of the next, or the "initiation," as the case may be.

ERUPTION PHASE

The eruption phase is that portion of the cycle during which the eruptive characteristics of the geyser are evident, from the moment of initiation to end of the falling stage, inclusive.

Initiation.—The exact time of the appearance of the first discharge of water above the mouth that is followed by the characteristic eruptive display. In most cases it will be followed by continuous discharge of water until the end of the eruption is reached.

Rising stage.—That portion of the phase between the initiation and the maximum stage.

Maximum stage.—That portion of the phase during which water is being thrown to the greatest height. In some geysers this will be only a moment, while in others it may be several minutes. In the latter case, there may even be several successive maxima, separated by discharge of lesser height. In such a case, the series should be called maximum stage and may be divided into serially numbered substages and pauses. The observer should consider the stage ended as soon as it is apparent the maximum height of the eruption will not again be attained.

Falling stage.—That portion of the phase between the end of the maximum stage and the disappearance of the last discharge of water above the mouth of the geyser. However, should an occasional spurt of water be thrown out as a result of water running back into the mouth and interfering with the steam discharge, it should be classed under the steam phase and not as part of the discharge proper.

STEAM PHASE

The portion of the cycle from the end of the falling stage to the beginning of the quiet phase constitutes the steam phase. The stages are as follows:

Violent steam stage.—That portion of the phase during which steam is emitted from the mouth in violent puffs or in a continuous column and is accompanied by distinctive sound.

Passive steam stage.—That portion of the phase during which emission of steam is at low velocity and is not accompanied by roaring or puffing sounds. The end may be an arbitrary limit.

QUIET PHASE

The quiet phase is that portion of the cycle between the end of the steam phase and the beginning of the next premonitory phase. It is not necessary to divide it; but, if for any reason division should be desired, the observer should employ suitably named stages in keeping with the other phases.

DEFINITIONS

Geyser mouth means that orifice through which is discharged the column of water reaching the maximum height.

Water column is the main column of water discharged by the geyser, whether vertical or not.

Steam column is the column of steam rising as the result of the eruption and does not include the water column. It may be employed, however, for steam rising during any phase of the cycle.

"Above the geyser mouth" is defined as "above the horizontal plane, passing through the lip." The "base of the water column" is the intersection of this plane with the water column.

Rim is the watershed around the mouth.

Lip is the lowest point of the rim. In certain geysers the second lowest point of the rim is also significant. This may be called the "second lip."

Maximum height of the water column should be taken as the vertical distance between the base of the column and the tip of the highest splash.

Maximum displacement of the water column should be taken as the distance between the base and the tip of the farthest splash from the column, whether the column is vertical or not. If the column is vertical, the displacement and the height will be the same.

Interval is that period of time between two successive initiations.

Frequency is the number of cycles per unit time, whatever unit may be employed for the particular geyser.

Pulsations is the term used for the number of discharges of water per second during the eruption phase.

Velocity of discharge is the distance traveled by a discharge of water, divided by the elapsed time.

SPECIAL CASES

It is suggested that all geysers be divided into three types, on the basis of their physical behavior.

Type I.—All geysers whose behavior can be described by the suggested nomenclature. It is understood that a geyser may be of this type even though it may not display all the phases and stages. For example, Daisy Geyser has no steam phase, yet it clearly belongs to this type.

Type II.—Those geysers whose cycles can be described by merely subdividing one or more phases into subphases. For example, the Grand Geyser's activity may be described adequately in this way, although it clearly cannot be classed as Type I.

Recording the eruption of the Grand would involve these changes: (1) division of the eruption phase into eruption subphases, which are serially numbered; (2) elimination of the term "initiation" from all subphases following the first; (3) considering the time between the end of the falling stage of one subphase and the beginning of the rising stage of the next subphase as the last stage of the former. It should be called the "negative stage" of the subphase.

Type III.—Those geysers whose behavior cannot be described easily under the first two types. The nomenclature may be modified as necessary.

TECHNIQUE

When data are given concerning geysers, it shall be understood such data are estimated unless specific details of instruments and methods used are stated.

For observation of time, where serious errors creep in due to inaccuracy of the chronometric device or due to discrepancy between two or more such devices used by one or more observers, it is suggested that no measurement of time should be called "accurate"

unless it is made by a watch or other device whose departure from correct time in a twenty-four-hour period is less than one minute.

If more than one timepiece is used, all readings should be reduced to standard, but the type and variation of each should be stated. If the departure from standard is known and stated for each timepiece used, correction can be made.

For linear measurements a steel tape should be used. If measurements are made from a map, an accurate chain scale should be employed and the map should be named in the record. Triangulation may also be used if an accurate instrument is employed. Any of these may be considered "accurate."

For angular measurements, no instrument of an order of accuracy less than that of the Brunton compass should be used if results are to be considered "accurate."

When an observer proposes to record a geyser eruption, he should prepare, in advance, sheets bearing the proper divisions. This will greatly facilitate the making of observations, for it then will be necessary only to fill in the proper blanks. Identification of the initiation will be easy, for it will be recorded as it comes and need not be recognized as such at the moment.

Time should be recorded to the second. The form of the record should be: "A.M. $10^h14^m11^s$ to $10^h18^m33^s$," and not merely as "4 minutes 22 seconds."

It is suggested that wind direction and velocity, air temperature, and other weather data be recorded. It is believed these may be of value. If it is apparent that the height of the water column is reduced by a strong wind, as is often true of Old Faithful, the fact should be recorded.

SUMMARY

It is understood that the observer need not make observations in such detail if he does not choose. The nomenclature is designed to provide a standard form for recording data. Adherence to this system by all observers will make it possible to correlate all data. Thus, if the observer records only the time of the initiations of two eruptions and makes no use of any other part of the nomenclature, the record will be useful.

The student of geyser behavior in past years should ascertain, if possible, the basis used by the various observers and should then convert such records into the terms of this nomenclature. If the basis cannot be ascertained with certainty, the data should not be used in compilations of "accurate" records, but should be separately mentioned and explained.

It is believed that only records made with great care and in detail will prove of much value in the future. The behavior of geysers which seem utterly capricious, if carefully recorded, may lead to better understanding. Hence, no geyser should be neglected.

REVIEWS

Physiography of Eastern United States. By NEVIN M. FENNEMAN. New York: McGraw-Hill Book Co., 1938. Pp. 691; pls. 7; figs. 197. $6.50.

In the companion volume to *Physiography of Western United States* (1931), Professor Fenneman has organized the available information on the geomorphology of the United States east of the Great Plains. Anyone who has attempted to keep abreast of present-day literature on the physiography of this vast area will realize the magnitude of this task and appreciate the immense amount of effort involved in the preparation of this book. Several improvements over the former volume are apparent, especially the inclusion of more geologic cross sections, block diagrams, and tables of formations. The discussion of controversial points is decidedly better although still rather noncommittal.

The discussion of the Coastal Plain makes available many new data and interpretations and fails only to mention the implications of the nonmarine Pliocene gravel mantle, which certainly once was much more extensive and might possibly be associated with pedimentation farther inland, especially in the Piedmont. In describing that province, the relation of the upland surface to the peneplain under the Cretaceous is fully explained, as well as Campbell's hypothesis of recent deformation of the Piedmont. The Crystalline Appalachians are named the Blue Ridge Province, a designation open to criticism, as the term Blue Ridge is commonly applied only to the escarpment along its eastern border. The discussion of the striking contrast between the Blue Ridge and the Piedmont is noncommittal and might well be fuller. The Folded or Newer Appalachians are named the Ridge and Valley Province and are discussed in great detail. Although Fenneman accepts lowering of the ridge crests during dissection of the ancient summit peneplain, he does not make clear the fact that the interpretations of ridge crests and rock terraces have ranged from a multitude of partial peneplains found by drawing straight or curved lines on cross sections to Ashley's suggestion of only one peneplain which has been differentially eroded with preservation of accordant ridge crests. One looks in vain for a full discussion of the old hypothesis that stairways of peneplain remnants survive without notably more dissection of the older than of the later terraces, that is, the problem of

lateral planation versus weathering, slopewash, and creep in peneplain formation. Meyerhoff and Olmsted's hypothesis is not mentioned. In describing the glaciation of part of this province Fenneman is probably the first physiographer to recognize the modern idea of stagnation of continental glaciers during dissipation.

In describing the Appalachian plateaus, regional peneplanation is recognized and the several hypotheses of age relations are taken up fully. Glacial alteration of the valleys in which lie the Finger Lakes is recognized by the statement that "nothing but a glacier could have made the sides of these lake basins so simple," but preference is given to stream work as the cause of the extreme depth of the lakes. The book was evidently completed before the publication of Ver Steeg's work on the subdrift topography of Ohio in which the "Deep Stage" of interglacial stream erosion was discriminated. The treatment of the New England Province is open to little criticism and includes both Barrell's hypothesis of marine terracing and Flint's interpretation of the Connecticut River terraces. The Adirondacks are considered as a separate province instead of as part of either the Laurentian or the Appalachian uplifts. The Interior Low Plateau Province embraces the Highland Rim with its included basins. The Lexington peneplain is correlated with the Highland Rim peneplain and the Nashville Basin with the narrow valleys below the Parker strath of the Appalachian Plateau.

The Central Lowland is divided into two portions, namely, east and west of the Mississippi River, and the Superior Upland is described between these because of its geographic location. Base-leveling of the nearly horizontal strata of the lowland is accepted because the amount of erosion above the crests of the cuestas varies so widely, but Martin's objections to the interpretation of the uplands as remnants of ancient peneplains is dismissed with the statement: "It should be unnecessary to argue the case for past peneplanation in a region where the real problem is how it could escape being peneplaned ten times over." The only serious omissions which came to the notice of the reviewer are, first, the fact that the major part of the peneplanation of the folded strata of the Arbuckle Mountains is demonstrably of Pennsylvanian age (pre-Pontotoc) and, second, the explanation of the Till Plains section as a consequence of the clayey nature of the drift. Little comment is needed for the treatment of either the Ozark or the Ouachita provinces.

The physiographic maps at the end of the book are an improvement over those of the former volume in so far as they show boundaries of provinces in a red overprint. However, in the expression of topography they are sadly inferior to the excellent diagrams by Guy-Harold Smith

which preceded them. In general, the reproduction of photographs is mediocre on account of the quality of paper. The reviewer must also note with regret the lack of a uniform system of organization and of clear-cut comparisons of different hypotheses. However, these omissions do not greatly detract from the value of the work, and beyond doubt all who are interested in physiography owe a debt of gratitude to Professor Fenneman for placing such a vast amount of information in a single volume.

F. T. THWAITES

Austrian Earth-Science in Sketches (*Oesterreich in erdkundlichen Zeichenskizzen*). By FRITZ STRADNER and FERDINAND SCHWARZ. Innsbruck: Tyrolia Verlag, 1937. Pp. 6; pls. 32. 6*s*.

This is a delightful visual account for use by the layman. Most of the thirty-two plates (really collections of simple line drawings) give a connected description of the geologic, physiographic, or cultural features that typify some province of Austria. For example, one page is devoted to the process of erosion by water and the resulting land forms. Another outlines the "Wald" part of lower Austria. As further illustration, the latter shows in separate small sketches or maps the drainage of the region, the outline geology of the bedrock, the origin of the scarp largely bounding this area, the structure (generalized) of the region, the relative altitudes of the smaller physiographic units, the meanders of a prominent stream, the control which valley form has exercised on agricultural development, the origin of one of the picturesque "Insel-Berge," and the relation between topography and situation of the more ancient castles. Some charts are more essentially geologic than that described, others less so.

The publication makes no pretense at offering scientific detail or explanation. But it furnishes the casual traveler with a glimpse of the outstanding geologic and geographic features of the countryside, stripped of the technical terms and the wordy descriptions that so generally intervene between science and the layman.

CHARLES H. BEHRE, JR.

"The West Nelson Earthquakes of 1929," by J. HENDERSON, in *Department of Scientific Industrial Research Bulletin 55* (extracted from *New Zealand Journal of Science and Technology*, Vol. XIX [1937], pp. 65–144). Wellington, New Zealand: Government Printer. 2*s*.

Reliable determinations of the absolute movement in faulting, as distinguished from mere relative displacement, are greatly needed. Such

information is now afforded by this study of the West Nelson earthquakes of 1929, owing to the fact that a railroad crosses the active faults and that two resurveys of the railroad levels were made following the main shock and the continuing after-shocks. Surveys of this sort before and after important earthquake faulting are just what have long been wanted to give us the nature of the faulting in the Great Basin of the western United States. For the Great Basin the crucial question has been: Have the mountain blocks risen or have the intervening basin blocks moved downward?

The general structure of West Nelson has resulted from horizontal compression which has produced a succession of high-angle thrust faults, many of which are in couples, outlining wedge-shaped upthrust blocks. The author believes that all the mountain-building faults in West Nelson are of the reverse type and that the mountain blocks are characteristically wedges tapering in depth. The chief displacements occurred in the Pliocene; those since the close of that period have been very much less.

During the earthquake of June 17, 1929, the principal surface of slip was the White Creek fault. Where the fault crossed the railroad "the road surface on both sides of a sharp single break appeared undisturbed except that that on the east side was 14 feet 9 inches higher than that on the west." A displaced fence not far away showed that the east block had moved, relative to the west block, 7.2 feet north and 5 feet west. The 5-foot west movement, combined with the 14 foot 9-inch uplift, is taken to indicate an east dip of 70° for the plane of slip.

The railroad route was releveled in August, 1929, and again during the March–July period of 1930. The first series of relevelings "showed that the country east of the fault had been raised and tilted, the uplift greatest near the White Creek fault, decreasing eastward, and disappearing about twelve miles east of the fault." The releveling indicated an uplift of 14.75 feet at the easterly dipping White Creek fault, increasing to a maximum of 16.08 feet about a mile farther east just before reaching a subordinate (Marina) fault. Between these two faults is a subsidiary wedge block whose eastern margin attained the greatest uplift of the whole elevated mass. "Levels west of the White Creek fault show uplift of 5 feet for a few chains, followed for a further distance of 15 chains by small uplifts decreasing rather irregularly westward and ranging from 15 inches to 1 inch." This apparently represents the drag on the relatively passive downthrow block.

For the most part the movements between the first and second releveling were downward, but the settling was uneven, suggesting differences

in behavior of several separate blocks. The greatest and most uniform settling occurred over a strip of 5 miles in the middle portion of the 12 mile section. This block, which had been raised and tilted evenly, sank as a unit about 18 inches during the period of some ten months between the reobservations.

The West Nelson earthquake seems to answer the question, sometimes raised, as to whether strong earthquakes are caused only by normal faulting or whether they may result also from thrust faulting. Also, since surveyed points west of the White Creek fault moved upward, west, and north in the same directions, though much less than surveyed points east of the fault, there seems to the reviewer to be no reason to appeal to elastic rebound as a factor of any consequence in producing the fault displacement. The faulting appears to have been, instead, a direct forward shove of an active block which rode obliquely upward above a relatively passive tract on the other side of the fault. Except for steeper angle and greater strike slip, it seems to have occurred much like the familiar thrust faulting experimentally produced in a pressure box.

In this mountainous region the earthquake produced innumerable landslips, rock slides, and rock falls, many of which were on a very large scale. These are described in detail in this *Bulletin*.

R. T. C.

The First Hundred Years of the Geological Survey of Great Britain. By Sir John S. Flett. London: H.M. Stationery Office, 1937. Pp. 280; pls. 13. 7*s*. 6*d*. ($2.15). Obtainable at the British Library of Information, 270 Madison Ave., New York City.

The Geological Survey of Great Britain celebrated its hundredth anniversary in 1935. The present volume is intended to commemorate that event as well as to review the century of useful and scientific service performed by the Survey. Consequently there are two parts, of which the first (196 pp.) is historical while the second describes the anniversary ceremonies.

The history of the Survey is divided, for convenience, into seven periods corresponding with the terms of its seven directors to date: De la Beche, Murchison, Ramsay, Geikie, Teall, Strahan, and, finally, Sir John Flett himself. An introductory chapter describes the circumstances leading up to the initial organization of the Survey under the leadership of De la Beche. The subsequent history is woven around the long series of projects which it has undertaken and tells how each was turned to achievement through the guiding efforts of that small group of

eminent scientists which has consistently formed the nucleus of the Survey. Changes in organization and technique incident to the advance of the science and the circumstances of the times are, of course, an integral part of the narrative, together with the inevitable changes in personnel through the years. While neither personalities, nor political undercurrents, nor the development of scientific concepts are accorded undue prominence, their presence as determining factors behind the activities of the Survey are felt with sufficient force to maintain the continuity and cohesion of the story.

Portraits of the various directors and other prominent members of the staff accompany the text. In addition, references to some seventy-five useful sources of biographical and historical material are listed in a Bibliography.

WILLIAM F. READ

Scientific Illustration. By JOHN L. RIDGWAY. Stanford University: Stanford University Press, 1938. Pp. 173; pls. 22; figs. 23. $4.00.

This book is designed to serve as a manual of correct procedure in the preparation of illustrations for scientific publications. Three principal types of illustrations are discussed: drawings, photographs, and maps. It is commonly believed that success in drawing requires a certain constitutional aptitude for artistic expression. Mr. Ridgway, however, points out that the only essential requirements, as far as scientific drawings are concerned, are an observant eye and a capacity for taking pains. The effectiveness of drawings as they finally appear in published form depends in large measure upon good judgment and foresight on the part of the illustrator in choosing appropriate subjects, in selecting the style of drawing best suited to his ends, in using proper drawing materials, in orienting and lighting his subject correctly, and in observing various necessary precautions in order that his drawings may not suffer in the processes of reproduction. In scientific photographs the important point is that they should clearly elucidate the point which the author wishes to make. In this connection Mr. Ridgway offers many valuable suggestions on special photographic techniques and methods of preparing photographs to display important features to best advantage. The methods of preparing various kinds of maps are considered in detail, with attention to such matters as the selection of base maps, use of symbols, techniques in lettering, etc. Structure sections, columnar sections, and block diagrams are discussed. The particular process of reproduction to which the illustrator's work will eventually be subjected is an important factor in

determining the method he shall use in preparing material; consequently, a section of this book is given to consideration of the various processes most commonly employed. Geologists and paleontologists will be pleased to note the special attention given to types of illustrations commonly employed in their fields of science. Mr. Ridgway was, for twenty-two years, in charge of illustrations for the United States Geological Survey.

WILLIAM F. READ

"Eocene Flora of Western America," by ETHEL I. SANBORN, SUSAN S. POTBURY, and HARRY D. MACGINITIE. *Carnegie Institution Washington Pub. 465.* Washington, D.C., 1937. Pp. 156; pls. 44.

Included in a single volume are three monographs dealing with early Tertiary floras of the west coast: Part I, "The Comstock Flora of West Central Oregon," by Ethel I. Sanborn; Part II, "The La Porte Flora of Plumas County, California," by Susan S. Potbury: Part III, "The Flora of the Weaverville Beds of Trinity County, California; with Descriptions of the Plant-bearing Beds," by Harry D. MacGinitie.

The Comstock flora, consisting of 739 specimens, was collected from two localities on the west side of the Cascades. The collection includes 29 species, of which 10 are new. The modern species which they resemble are indigenous to warm latitudes. The age of the flora is Middle Eocene, at which time a temperate to subtropical climate is indicated for the region.

The La Porte flora occurs in a dacite tuff overlying carbonaceous shales and auriferous gravels in Plumas County, California. Forty-one species are recognized, of which 27 are new. All have living equivalents inhabiting warm temperate to tropical environments. Fossil and modern relationships indicate that the La Porte flora is either Eocene or early Oligocene in age. Stratigraphic relationships tend to indicate that it is not older than Upper Eocene.

The Weaverville flora consists of 37 species, 17 being new. The specimens were collected from two localities in Trinity County, California, where they occur in beds of fluviatile and lacustrine origin. Living equivalents are largely confined to humid, warm temperate regions of Asia and North America. The flora is provincial in character, with no absolute conclusion being reached as to its exact age. Comparison with other Tertiary floras indicates a possible range from Middle Eocene to Upper Miocene, with the probability that it is of older Oligocene age.

The three monographs are uniform in style and include descriptions

of all forms occurring in the respective floras. The volume is profusely illustrated with unusually well-executed photographs.

RAYMOND E. JANSSEN

Quartz Family Minerals. By H. C. DAKE, FRANK L. FLEENER, and BEN HUR WILSON. New York and London: McGraw-Hill Book Co., 1938. Pp. xvi+304. $2.50.

The subtitle of this book, *A Handbook for the Mineral Collector,* expresses adequately the purpose for which it is written. The authors evidently understand clearly the enthusiasm with which the amateur collector of minerals pursues his hobby. They are aware also of his need for ready information about the common mineral, quartz. To meet his need they have elaborated fully, and yet in nontechnical style, a wide range of topics concerning quartz: its occurrence, its varieties, its unusual types, and, the subject so fascinating to the mineral collector, silicified woods.

Throughout the book the authors consider the status of the lay reader and do not affright him with the tone of a textbook. The work, accordingly, is of intrinsic value to the amateur collector for its information, for its suggestions on collecting and preparing specimens, and for the emphasis it gives to the scientific interest bound up in the quartz family.

An appendix of selected references is added for further reading about the physical and chemical properties of the minerals.

JOHN R. BALL

The Laws of Rock Metamorphism. By V. M. GOLDSCHMIDT (trans. TERENCE T. QUIRKE). Champaign, Ill.: University of Illinois Supply Store, 1937. Pp. i+28. $0.65.

This is a mimeographed edition of a translation into English of two papers by Professor Goldschmidt published some twenty-five years ago.

The first is the paper in which Goldschmidt discussed, as a temperature-pressure indicator, the reaction between calcium carbonate and silica to give wollastonite and carbon dioxide, and in which he developed from thermodynamic consideration his well-known curve for this equilibrium.

The second is one of the series of papers on the rocks of southern Norway, in which field and microscopic methods are combined to attack problems of regional and contact metamorphism and enable a determination of the factors that control the contrasted results.

The student may find it useful to have these original sources available in translation.

N. L. B.